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SOUND VELOCITY, ELASTICITY, AND RELATED PROPERTIES OF MARINE SEDIMENTS, NORTH PACIFIC

III. PREDICTION OF *IN SITU* PROPERTIES

E. L. HAMILTON

OCEAN SCIENCES DEPARTMENT SAN DIEGO, CALIFORNIA OCTOBER 1969

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THE PROBLEM

Determine and study the acoustic and related properties of the sea floor; specifically, predict *in situ* values of sound velocity, density, and other mass physical properties of sea floor sediments, and apply these predictions to studies of acoustic models of the sea floor.

RESULTS

1. This is the third of a three-part report on sound velocity, elasticity, and related physical properties of marine sediments, North Pacific; this report concerns the correction of laboratory to *in situ* values, and the prediction of *in situ* values. The methods are applicable to other areas and sediments.

2. The sediments studied can be assigned to three general sedimentary environments which are associated with three great physiographic provinces: (1) continental terrace (shelf and slope), (2) deep-water, abyssal plains (turbidites), and (3) abyssal hill (pelagic). The sediment types and their properties in a given area of the latter two environments can be predicted with some confidence; detailed sediment charts are required for the continental-shelf area.

3. The keys to predictions of *in situ* properties of marine sediments are knowledge of:

- a. general physiographic provinces and their sedimentary environments,
- b. sedimentary processes resulting in certain sediment types within these environments,
- c. laboratory or *in situ* values of the mass physical properties of the sediment types within the environments, and
- d. methods of correcting laboratory to *in situ* values.

4. The following sediment properties are reported.

- a. Sound (compressional-wave) velocity.
 - (1) Sound velocity, in deep-sea areas, can be predicted within about 1 to 2 percent for the sediment surface without any other sediment data (such as porosity, density, or grain size). Velocity in deep-sea areas should be predicted directly rather than predicting porosity or some other property from which velocity can be predicted.
 - (2) Predictions of velocity in continental-shelf areas are dependent on detailed sediment charts, or some sediment data.
 - (3) Two methods of correcting laboratory to *in situ* velocity in the sea floor are:
 - (a) Correct a laboratory value by applying temperature and pressure corrections from laboratory to *in situ* conditions, using tables for the speed of sound in seawater, or

- (b) The recommended procedure is to multiply the ratio, (velocity in sediment)/(velocity in seawater), by the velocity in bottom water in the area of interest; this ratio is the same in the laboratory as *in situ*; bottom-water velocities can be easily determined from a velocity vs. water-depth diagram which is valid over wide areas; although the numerical values of velocity in the surface sediment vary with the velocity in bottom water, the ratio remains the same.
 - (4) If no sediment data are available for a particular locality, an average velocity ratio for the environment and expectable sediment type can be taken from the tables and used for prediction of sediment-surface velocity.
- b. Porosity and density.
 - (1) Sediment porosity in the laboratory can be considered as the same *in situ*. True porosity can be computed from apparent porosity by making a correction to allow for the amount of dried salt weighed with dried mineral residues during laboratory procedures; this increment varies from 0.5 percent in sands to about 1 percent in high-porosity silt-clays.
 - (2) Density *in situ* is slightly greater than in the laboratory (heavier water in pore spaces). The correction, at a depth of 6000 m, amounts to about 0.025 g/cc, and can usually be disregarded as within the margins of error in laboratory procedures, or insignificant because of lateral and vertical changes in the sea floor over small areas and depths.
 - (3) Empirical corrections are furnished in tables for the above corrections to porosity and density; they can be applied by inspection.
- c. Impedance, or density \times (velocity).
In situ values of impedance should be computed after determination of *in situ* values of density and velocity.
- d. Rayleigh reflection coefficients and bottom loss at normal incidence.
 Laboratory values of Rayleigh reflection coefficients and bottom losses at normal incidence can be used as *in situ* values because of small differences. These computed values are close to those measured at sea.
- e. Elastic constants.
 Using measured values of density and velocity (corrected to *in situ* values), plus a computed value of the bulk modulus, the other elastic constants can be computed; these elastic constants include compressibility, rigidity (shear) modulus, Lamé's constant, Poisson's ratio, and velocity of the shear wave. Numerical examples are given of these computations for shallow and deep water.

5. Sound is reflected and refracted by layers within soft sediments, and by rock layers underlying the sediments. Whether or not a specific layer reflects sound energy depends on the frequency of the sound wave and properties of the layer. Density and velocity values are given for the principal types of layers in the sea floor; these are terrigenous sediments which may be lithified (or cemented) to form mudstone and shale, calcareous materials which may form limestone, siliceous sediments which may form various types of siliceous rocks (including diatomite and chert), and layering due to volcanic ash and basalt.

- a. The igneous rock, basalt, forms the upper oceanic crust of the earth, but is also at the sea floor, or just beneath soft sediments over wide areas; consequently, the properties of basalt can be of importance in underwater acoustics. A special section is devoted to a discussion of velocity and density values expectable in oceanic basalts, and a method is proposed for computing density when velocity is given (as from a seismic refraction survey).
6. Velocity gradients within sediment bodies.
- a. Variations of velocity with depth in sediments are required to compute true thicknesses of layers, and for certain studies in underwater acoustics. Velocity gradients usually vary between 0.5 and 2.0 sec⁻¹ at the sediment surface and can be considered linear to small depths. Curves of velocity vs. depth or time are parabolic to greater depths, and gradients decrease with increasing depth in thicker sediments.
 - b. The present method of determining interval velocities in layers, variations of velocity with depth or travel time, and velocity gradients, is with expendable sonobuoys. This technique has yielded much information on these properties in the Atlantic and Pacific. Values determined by the Lamont-Doherty group are in good agreement with those determined at this Center and at Scripps Institution (reports in preparation); curves, equations, and values are given in a resumé of the Lamont-Doherty data, and recommendations are made for their use in the Pacific.

RECOMMENDATIONS

1. To predict *in situ* values of the mass physical properties of marine sediments, layers within sediments, and geoacoustic models of the sea floor for given areas, continue studies and surveys over a wide spectrum. The four most important are as follows:

- a. Studies of the topography, structure, and present sediment distributions of the sea floor; these allow separation of the sea floor into physiographic provinces and associated environments in which certain sediment types are found.
- b. Studies of sedimentary processes which indicate the method and route by which present-day sediments arrived at sites of deposition, and indicate the areal extent of these sediment types within the environments of a., above. Such studies allow extrapolation, in many cases, of data from relatively few sampling sites.
- c. Studies of the location and thicknesses of reflecting layers, at depth within sediments; these can be determined over large areas, only by reflection-profiling. Reflection-profiling also guides sampling operations (where the reflecting layers crop out on the sea floor).
- d. Studies of the mass physical properties of sediments from wide areas; these should continue to determine the parameters and statistical variations of

these properties. Both laboratory and *in situ* measurements should be made to determine suitable corrections, and to improve present sampling techniques.

2. Special attention should be drawn to the following recommendations:

- a. Strongly support present efforts of the Naval Oceanographic Office to collect data and publish atlases (and revise old atlases) which give information on the water masses and sea floor.
- b. Encourage deep-sea drilling (such as that of the present JOIDES program); such efforts are necessary to identify and determine properties of reflecting layers in the sea floor below coring depths.
- c. Accelerate the study of two sediment properties which are important in understanding the reflection of sound from the sea floor: shear-wave velocity, and attenuation of compressional waves at various frequencies of interest in underwater acoustics and geophysics.

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PREFACE

This report is Part III in a series of Technical Publications on sound velocity, elasticity, and related properties of marine sediments from three major environments of the North Pacific: the continental terrace (shelf and slope), abyssal plain (turbidite), and abyssal hill (pelagic).

Part I (TP 143) detailed the measurement and computation of mass physical properties of the sediments, including density, velocity, grain size, impedance, reflection coefficients, and bottom loss.

Part II (TP 144) presented discussions of elastic and viscoelastic models for water-saturated porous media, and measurements and computation of elastic constants.

The present report (TP 145) is concerned with predictions and computations of *in situ* physical properties.

Throughout each report, references are made to the other two studies, as appropriate.

INTRODUCTION

This is the third of a three-part report on the methods, techniques, measurements, and computations of mass physical properties of marine sediments from several major environments in the North Pacific and adjacent areas. Within each of these environments, data are given for each of the major sediment types. Calcareous ooze is the only major sediment type not represented in the present study.

In Part I, the following properties and their interrelationships were listed and discussed: mean and median grain diameters, percentages of sand, silt, and clay, saturated bulk densities, bulk densities of mineral grains, porosities, sound velocities, impedances, density \times (sound velocity)², the ratios: (sound velocity in sediment)/(sound velocity in seawater), and Rayleigh reflection coefficients and bottom losses at normal incidence. The data were presented in tables, and their interrelationships were shown in scatter diagrams and regressions equations.

Part II was concerned with the elastic properties of the sediments, including elastic models, and measurements and computations of elastic constants and their interrelationships with other physical properties. These discussions included compressibility, bulk modulus, rigidity (shear) modulus, Lamé's constant, Poisson's ratio, density, and shear- and compressional-wave velocities.

This part of the study is concerned with the predictions of *in situ*, sea floor values of those sediment properties discussed in the first two parts. Although the sediment properties, examples of computations, and tables of this report are related to the North Pacific and adjacent areas, the methods and techniques of prediction, and correction of laboratory values to *in situ* conditions are applicable to other areas and sediments.

In both Parts I and II the values of all properties were for the laboratory at 1 atmosphere pressure; where temperature was an important factor, the measurements were corrected, or referred to 23°C.

The geophysicist, geologist, and acoustician specializing in underwater sound require values of the acoustic properties *in situ*, and not in the laboratory. There has been a tendency to assume that values of sound velocity (compressional-wave velocity) measured at 1 atmosphere, and any "room temperature" can be used for *in situ* values because temperature and pressure corrections (laboratory to sea floor) will cancel out. This is true only for specific conditions in the laboratory and the sea floor; otherwise, errors on the order of 1 to 5 percent of sound velocity are expectable. There are also changes in density and elastic properties; consequently, any computations involving sound velocity, elasticity, and related properties of sediments on the sea floor should use *in situ* values.

The empirical relationships between physical properties was discussed in Part I, and basic, theoretical relationships reported in Part II. This part will be confined to practical, step-by-step methods of estimating and computing *in situ* properties of marine sediments. Each step will be illustrated by numerical examples.

PHYSIOGRAPHIC PROVINCES AND ASSOCIATED SEDIMENTS

In the North Pacific and adjacent seas there are, in general, the following great physiographic provinces (figure 1):

1. Part of the *Pacific Basin*, which contains
 - a. *abyssal hills* which form about 80 percent of the bottom topography of the Pacific Basin.

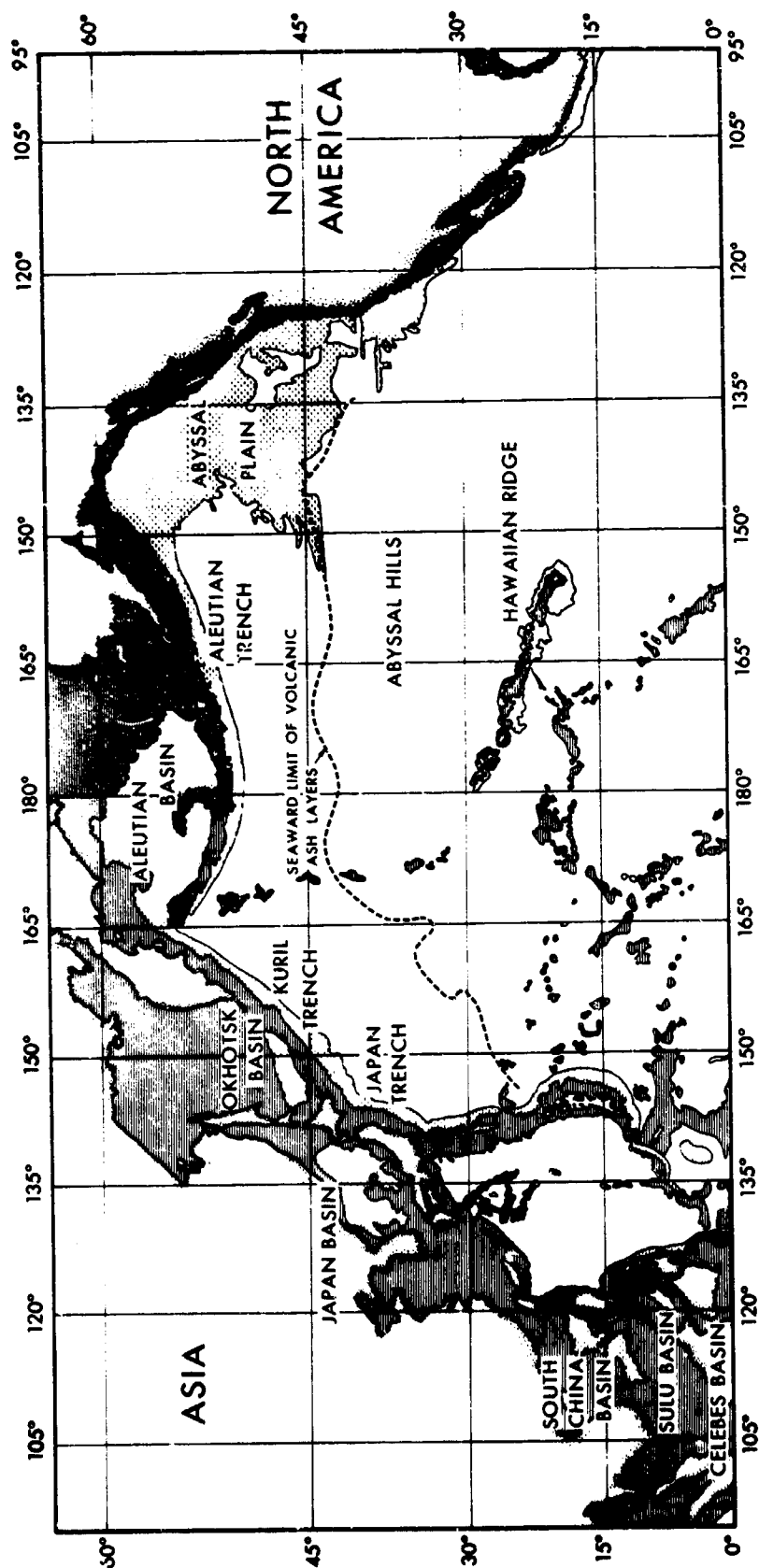


Figure 1. Physiographic provinces and related sedimentary environments, North Pacific and adjacent areas; seaward limit of volcanic ash layers from Horn *et al.* (1969a).

- b. *large rises and ridges, and chains of islands and seamounts* (e.g., the Hawaiian Ridge, Emperor Seamount Chain, Marcus-Necker Rise, East Pacific Rise), which are surrounded, in many cases, by
- c. aprons of sediment (*archipelagic aprons*), and
- d. *abyssal plains* which are confined to the northeast part of the North Pacific Basin adjacent to the North American continent, (e.g., the Alaskan and Tufts Abyssal Plains in the Gulf of Alaska).

2. Peripheral to the northern part of the Pacific Basin are

- a. the *great trenches* which border much of the basin (e.g., Aleutian Trench, Kuril-Kamchatka Trench, Japan Trench),
- b. the *deep basins* ("peripheral basins") of the seas *adjacent to the Pacific Ocean* (e.g., Aleutian Basin, Okhotsk Basin, Japan Basin), and
- c. the *continental terrace* (continental shelf and slope).

Relatively minor features include the larger seamounts, areas within fracture zones, and other features which, in generalizing, must be included in larger provinces, especially when sediment types and properties are the subject of study.

Because this report is concerned with the properties of sediments in the North Pacific, the general subject of the physiography of the area will not be elaborated; see Menard (1964) and Horn *et al.* (1969) for general discussions.

Each of the physiographic provinces noted above forms sedimentary environments in which distinctive sediment types are apt to be found (the nomenclature and principal types of these sediments are discussed in Part I). The processes involved in forming these distinctive sediments are complex and are the subject of a vast literature. The construction of sediment charts, and the prediction of sediment types and their properties in any environment, are dependent on studies of sedimentary processes; consequently such studies form an important part of the information on which the prediction of any property is based. The statements above are inserted to answer a question sometimes posed in the field of applied oceanography: what is the connection between basic research studies of sedimentary processes, and the acoustic properties of sediments?

Sediment sampling in the North Pacific and adjacent areas is now sufficient to permit generalized charts of surface sediment distributions for the whole area. Recent examples are by Bezrukov (1960), Shepard (1963), Arrhenius (1963), Menard (1964), Keller (1968), and Horn *et al.* (1968, 1969). Newer charts are being prepared by Arrhenius at Scripps Institution, and in atlases by the U.S. Naval Oceanographic Office. Detailed charts are available in many smaller areas.

On continental shelves, the sediment types range from sands and gravels to silty clays; they include a full range of mixtures of these end types. The placement of these sediments on continental shelves includes all sedimentary processes: transportation and reworking of mineral and biogenous particles by waves and currents, deposition of individual particles, vertically, through the overlying water column ("pelagic sedimentation"), sediments deposited from turbidity currents moving along the sea floor to form "turbidites," slumping and sliding on the steeper slopes, and other processes. The types of sediments on the continental shelf are apt to change drastically over small areas, both horizontally and vertically; consequently, prediction, in detail, of a sediment type within a given locality is heavily dependent on actual sampling or on detailed, accurate sediment charts (for surface sediment types), and on acoustic reflection or borings (to determine the thicknesses of surficial and buried layers).

The sediments on continental slopes are dominantly mixtures of silt-clay sizes. In the samples of this study no distinction could be made between the important mass physical properties of these fine-grained sediments and those of the same type on the continental shelf; consequently, sediments from the shelf and slope were placed in the same environmental category: "Continental Terrace (Shelf and Slope)." The few island shelf and slope samples were placed in this "Terrace" category. It is probable that future studies will indicate that the sands, silts, and clays of the narrow shelves and slopes of islands can be placed in the Terrace category (which could then be renamed "Continental and Island: Shelf and Slope").

Sediments deposited from turbidity currents (turbidites) cover rough topography and form flat areas of the sea floor. Turbidites are normally expectable in basins within the continental terrace, in the flat floors of the peripheral basins (*i.e.*, Aleutian, Okhotsk, and other basins), the great trenches, and in deep-water abyssal plains. Turbidity currents, apparently, have not been active in the peripheral basins and deep abyssal plains since the Pleistocene, because most of these areas are covered by a thin (a few centimeters) layer of "red" clay. Below the sediment surface in these areas, thin lenses and layers of sand, silty sand, sandy silt, and coarse silt are usually intercalated between thicker layers of greenish-gray, higher-porosity silty clay, clayey silt, and clay. Layers of volcanic ash are common in sediments adjacent to volcanic areas. Prediction of sediment types and their properties is, however, usually easier than on the shelf; especially in the floors of abyssal plains which are distant from land. Predictions of properties of turbidites in basins in the continental terrace (*e.g.*, the San Diego Trough), or in abyssal plains close to continental areas (*e.g.*, the Tyrrhenian Abyssal Plain: Kermabon *et al.*, 1969), can be as difficult as for shelf sediments. In both cases the difficulty is due to the vertical and lateral variations in sediment types and properties.

In this report, the deep-water turbidites are placed in the same environmental category, "Deep-Water, Abyssal Plain (Turbidite)," which includes samples from the abyssal plains of the peripheral basins (including the Aleutian, Okhotsk, Japan, South China, Celebes, and Sulu Basins), and the floors of the deep trenches around the margins of the Pacific Basin (including the Japan and Middle America (Mexico) Trenches).

The category "Deep-water, Abyssal Plain (Turbidite) Environment," as used in this paper, requires some discussion. All of the samples from this environment are either from deep trench floors or from the central parts of the deep, flat basins of the seas peripheral to the Pacific Basin -- for example, the Aleutian Basin (Bering Sea), and the Japan Basin (Japan Sea). In these areas, the sediment surface is usually a high-porosity, fine-grained silt-clay, overlying buried layers of sand-silt (including volcanic ash). Because the samples reported herein are from the upper 30 cm, these coarser-grained, higher-velocity layers are not adequately represented (statistically). The properties of these layers will be the subject of a special section. However, the number of high-porosity samples allows comparisons with the high-porosity sediments from the other environments (see Parts I and II).

In the abyssal-hill environment, the common sediment types are few, and are predictable over very wide areas. In the North Pacific, the common sediment type is silty clay or clay (deep-sea, "red" clay). After comparison, samples from distal ends of aprons around ancient and modern islands ("archipelagic aprons") were included in the "Abyssal-Hill (Pelagic)" environment.

In summary, three general sedimentary environments (or provinces) are included in this report:

1. *Continental Terrace (Shelf and Slope)*, including the shelves and slopes around islands.

2. *Deep-Water, Abyssal Plain (Turbidite)*, abbreviated to "Abyssal Plain (Turbidite)," including the floors of the basins of the seas adjacent to the Pacific Basin, the floors of the deep trenches, and the deep, abyssal plains in the Pacific Basin. The shelves and slopes around the flat floors of the peripheral basins fall into category 1.

3. *Abyssal Hill (Pelagic)*, including the silt-clays ("red clays") of the abyssal hills and the distal ends of archipelagic aprons.

To make this part more useful, the tables from Parts I and II are included as Tables A-1 to A-4.* The sediment properties reported in these tables are from the sediment "surface" (0 to 30 cm).

It should be noted that the sediment samples, measurements, and discussions in Parts I and II did not include calcareous "ooze", which is not a common sediment in the North Pacific (north of the equatorial, calcareous area). Reports by Horn, *et al.* (1968a) and Schreiber (1968a) included calcareous sediments from the Atlantic and adjacent seas. Table 8 lists some representative properties of calcareous sediments.

The general surface sediment types and their properties in a given area can be predicted with some confidence. However, it is important to realize that below the sediment surface, in many areas, there are layers of volcanic ash and sand-silt of higher density and velocity which may be of more importance in determining reflection characteristics and interval velocities than the (usually) softer, overlying materials. In the absence of surveys and sampling, the detailed thicknesses, locations, and properties of these various layers within the sediment body cannot, at this time, be predicted with assurance. It can, however, be predicted that layering from deposition of volcanic materials will be present in some areas, and that layers of silt, or even sand, will be present in an abyssal plain or trench floor, deposited by turbidity currents. For the North Pacific, those areas in which such layering can be expected are illustrated in figure 1. The presence and extent of these layers is based on coring data, on acoustic-reflection surveys which revealed the actual presence of layering, and on a knowledge of sedimentary processes which allow the extrapolation of these data within the various physiographic provinces. The data of figure 1 were derived from the following sources (mostly, recent reports with numerous references to earlier work):

1. North Pacific Basin and surrounding trenches: Horn *et al.* (1969a,b); based on core analysis from Lamont-Doherty Geological Observatory.
2. Pacific Basin and adjacent areas: Menard (1964).
3. Aleutian Basin (Bering Sea): Ewing *et al.* (1965), Scholl *et al.* (1968).
4. Okhotsk Basin and adjacent areas: Bezrukov (1960).
5. Japan Basin: Kovylin *et al.* (1966), Hilde *et al.* (1969).
6. South China Basin: Ludwig *et al.* (1967).
7. Sulu and Celebes Basins: Krause (1966).
8. Trenches: Hurley (1960), Fisher (1961), Shor (1964), Ross and Shor (1965), Ludwig *et al.* (1966).

*Throughout the report, table numbers preceded by the letter A refer to tables in the Appendix.

9. Abyssal Plains, Gulf of Alaska: Hurley (1960), Nayudu and Enbysk (1964), Hamilton (1967), McManus (1967).

10. Hawaiian area: Hamilton (1957), Moberley and McCoy (1966), Schreiber (1968b), Lair and Sanko (1968).

11. Unpublished acoustic reflection data from Scripps Institution and the Naval Undersea Research and Development Center (formerly Navy Electronics Laboratory).

PREDICTION OF *IN SITU* MASS PHYSICAL PROPERTIES

Introduction

There is now enough information on sea-floor sediments in the various larger environments to allow, in the absence of specific data, reasonable predictions of several properties of importance in geophysical and acoustic studies. The keys to predictions of *in situ* properties of marine sediments are knowledge of:

1. general physiographic provinces and their sedimentary environments,
2. sedimentary processes resulting in certain sediment types (horizontally and vertically) within these environments,
3. laboratory or *in situ* values of the mass physical properties of the sediment types within the environments, and
4. methods of correcting laboratory to *in situ* values.

In the absence of specific sediment data for a given locality, the general method of prediction is as follows:

1. Predict the physiographic province with the aid of physiographic province charts, such as that of figure 1, or those of Horn *et al.* (1969), Menard (1964), Heezen *et al.* (1959), or the atlases of the Naval Oceanographic Office (e.g., Publication No. 700, Section V - N. Atlantic); physiographic provinces can also be deduced from bathymetric charts.
2. Predict the sedimentary environment and sediment type; a sediment chart should be used when available (thus obviating step 1 above); such charts employ sediment samples, and extrapolation of the sampled types, areally, within the sedimentary environments formed by the physiographic provinces; this type of extrapolation is based on a knowledge of sedimentary processes.
3. Predict the laboratory value of the desired property; sound velocity should be predicted directly rather than predicting density, porosity, or other property, and then using this property as an index to velocity, and
4. Correct the laboratory value to *in situ* conditions.

The preceding method, or sequence of predictions, reflects the present state-of-the-art. The measurement of sediment mass physical properties *in situ* from deep-diving submersibles, by divers, by remotely-controlled vehicles and devices, or by remote means is well underway, but, at present, is not sufficient to obviate the use of laboratory data. In the future (several years from now) charts of many areas of interest will display mass properties of the sediment body (horizontally and vertically) which were measured *in situ*. These charts will still be based on items 1 and 2 above (extrapolation of known data within physiographic provinces and sediment types), but for some properties it will not be necessary to use corrected laboratory values.

When no sediment-property data are available for a specific locality (3, above), the laboratory, average, value of the desired property should be taken from an appropriate table for the estimated province or environment, and sediment type. Tables A-1 and A-2 are reproduced from Part I of this report; these tables are summaries of the laboratory, average values for various mass physical properties of sea-floor sediments for the general environments of the North Pacific; both shallow and deep-water sediments are included. The tables can be used as bases for predicting these values, *in situ*, for given environments and sediment types in the North Pacific when no data are otherwise available. Similar data (in the form of diagrams) for the North Atlantic and adjacent areas can be found in Horn *et al.* (1968a) and Schreiber (1968a); and for parts of the deep Pacific in Horn *et al.* (1968e), and Schreiber (1968b).

When one is given sediment samples, or reports which list mass physical properties for a given locality, the available data can be used to enter empirical diagrams and equations which interrelate various mass properties and to derive laboratory values for those properties not measured or listed. In Part I of this report the empirical interrelationships between mass physical properties were presented in three forms: (1) scatter diagrams illustrating the relationships between two given properties, (2) regression equations (Part I, Appendix B) for the illustrated data, and (3) in tables listing the arithmetic mean (or average) and standard errors of the mean, for the properties of each sediment type within the three larger environments. Appendix A, Part I, includes discussions of the best empirical relationships and best entries into the data to arrive at various sediment properties. Additional data of a similar nature, for the deep North Pacific, are in Horn *et al.* (1968e) and Schreiber (1968b).

The use of laboratory values of sediment properties to get predicted *in situ* values is the subject of the remainder of this report.

Some of the mass physical properties of surficial sediments (0 to 30 cm depth) can be assumed to be the same in the laboratory as they are *in situ*; in this class are grain-size analyses (e.g., mean grain diameter, percentages of sand, silt, and clay sizes), porosity (salt-free), bulk density of mineral grains, and the ratio: velocity in sediment/velocity in seawater (to be discussed in detail later). Sufficiently different to require corrections from laboratory to *in situ* are saturated bulk density of the mineral-water system, sediment sound velocity (compressional wave), impedance (product of density and velocity), density \times (velocity)², and all of the elastic properties. Approximately the same, and sufficiently close so that laboratory and *in situ* values can be used interchangeably, are Rayleigh reflection coefficients and bottom losses (dB) at normal incidence (which were computed and discussed in Part I). Consequently, the sections below will include only those properties requiring corrections.

In addition to laboratory sediment properties, information is required before *in situ* sediment properties can be predicted; these are: sound velocity, density, and salinity of the seawater at the water-sediment interface. Useful, but not required if these water properties are known, are temperature and pressure of the water just above the sea floor. If a particular study requires sediment sound velocities over an area where topographic relief is appreciable, the investigator should have a table or profile of sound velocity vs. depth in the water column (e.g., table 1, figure 2), and a contoured chart of the

sea floor; information is needed, also, to convert echo-sounder depths to true depths. The sources of this information and its uses to derive properties of seawater will be discussed in connection with those sediment properties requiring it for *in situ* corrections.

TABLE 1. VARIATIONS OF PROPERTIES OF SEAWATER WITH DEPTH, CENTRAL PACIFIC*

Depth (m)	Temp (°C)	Salinity (ppt)	Pressure (kg/cm ²)	Sound Speed (m/sec)	Density (g/cc)
2000	2.0	34.60	207.61	1491.1	1.03708
3000	1.7	34.67	311.60	1507.0	1.04174
3500	1.5	34.68	363.76	1515.1	1.04402
4000	1.5	34.68	416.04	1523.9	1.04625
4500	1.5	34.68	468.43	1532.7	1.04847
5000	1.5	34.69	520.93	1541.7	1.05069
5500	1.5	34.69	573.54	1550.7	1.05287
6000	1.5	34.69	626.26	1559.8	1.05504

*Notes:

1. Temperature from Defant (1961; also in NAVOCEANO SP-68, Table 7) and Barbee (1965).
2. Salinity from Barbee (1965).
3. Pressure from Wilson (1959) for 35.0 ppt, 0°C.
4. Sound speed: entering tables for speed of sound in seawater (NAVOCEANO SP-58 or SP-68) with temperature, salinity, and pressure data indicated.
5. Density ($\rho_{s,t,p}$): reciprocal of specific volume, *in situ*; computed from H.O. 614 (LaFond, 1951; same tables in NAVOCEANO SP-68).

Sound Velocity

When a sediment sample is removed from the surface of the sea floor to the laboratory, the salinity of water within the pore spaces remains the same, and the only changes (disregarding possible disturbances to sediment structure) are in temperature and pressure. Loughton (1954, 1957) demonstrated, by laboratory tests, that sound-velocity changes in saturated sediment due to hydrostatic pressure were about the same as in seawater. Shumway (1958) demonstrated that velocity changes in saturated sediment (artificial and natural marine) due to temperature changes were about the same as in seawater. Hamilton (1963) measured sound velocity *in situ* from a deep submersible and in the laboratory from cores taken at the site of the *in situ* measurements and concluded that laboratory measurements could be corrected to *in situ* by applying full corrections for temperature and pressure in the laboratory to the temperature and pressure *in situ*, using tables for the speed of sound in seawater.

These findings allow two approaches in correcting laboratory values of sediment sound velocity to sea-floor values: (1) correct the laboratory value to *in situ* conditions by applying full corrections for temperature and pressure, using tables for the speed of sound in seawater (NAVOCEANO SP 58 or 68), or (2) use the ratio: sound velocity in sediment/sound velocity in seawater. The latter method is better. To understand the utility of this ratio, imagine that a unit volume of water adjacent to the sea floor, and a unit volume of water-saturated sediment from just below the water-sediment interface, are transported from the sea floor to the laboratory. The salinity of the bottom water is about the same as that within the pore spaces of the sediment (see discussion, Part I), and does not change from *in situ* to laboratory; therefore, the only changes in sound velocity in the bottom water and pore water are due to temperature and pressure changes. The effects of pressure and temperature changes on velocity in the minerals in the sediment are insignificant. This means that both water and sediment velocities can be corrected from laboratory to *in situ* by making the same corrections for temperature and pressure which are derived from tables for the speed of sound in seawater, and that *the ratio will be the same in the laboratory as in situ*; this was proved experimentally by measurements made from the bathyscaph TRIESTE (Hamilton, 1963) and the submersible, DEEPSTAR (partly in Hamilton *et al.*, 1969).

All the samples of this report were corrected to 23°C by using the Tables for the Speed of Sound in Sea Water (NAVOCEANO SP-58, 1962); the pressure in the laboratory is 1 atmosphere. Therefore, to obtain the ratio, sound velocity in the sediment/sound velocity in seawater, one need only divide the laboratory sediment velocity by the sound speed in seawater at 23°C, 1 atmosphere, and the appropriate salinity. The only variable in the seawater speed is salinity.

There is little variation in the salinity of bottom water in the deep, open oceans. For example, in the deep Central Pacific, bottom-water salinities generally vary between 34.65 and 34.70 ppt with an average value of about 34.69 ppt (Defant, 1961; Barbee, 1965). Bottom-water salinities off San Diego are about 33.5 ppt at 10 m, and about 34.5 ppt in the San Diego Trough at 1000 m. Given the same temperature and pressure, these small variations in bottom-water salinity result in very small variations in sound speed in the bottom water. Table 2 lists sound speeds and densities of seawater at 23°C, 1 atmosphere pressure, and various salinities. The small variations of sound velocity in seawater due to salinity variations make computations of the ratio (velocity in sediment/velocity in seawater) simple and valid, even when the salinity of bottom-water, or pore water, is not known for a specific sample, and an estimate must be used which has been derived from tabulated data (as in table 2), atlases, or textbooks. If the sediment velocity is measured at, or referred to, other than 23°C, the seawater sound speed can be computed from tables for 1 atmosphere pressure, the selected salinity, and the appropriate temperature. For sediments from the deep Central Pacific Basin, (at 23°C), the velocity value for 34.69 ppt (table 2) can be used with negligible error.

Scatter diagrams of the ratio vs. various other properties can be used to determine the ratio, but the best way to obtain this useful property is to determine the sediment sound velocity in the laboratory at a known temperature by measurement, or by entry into diagrams or equations and divide by the appropriate speed for seawater.

Having determined a value for the ratio, the *in situ* velocity in the surface of the sea floor is the product of the ratio and the bottom-water velocity. This "ratio method" is best because over any larger area of the sea floor the sediment velocity will change with the bottom-water velocity (which changes because of pressure and temperature effects, as water depth varies). It is simple, easy, and accurate to pick off the bottom-water velocity from a velocity-depth curve for the water mass at any given, true, water depth. A single velocity-depth profile, established by a Nansen cast or other means, will establish velocity-depth relations over very great areas for depths too great to be affected by seasonal changes in surface waters. For example, a single such profile in midlatitudes in

TABLE 2. DENSITY AND SOUND SPEED IN SEAWATER AT VARIOUS SALINITIES, AND AT 23°C. AND 1 ATMOSPHERE PRESSURE

Salinity (ppt)	Density (g/cc)	Sound Speed (m/sec)
33.00	1.02244	1527.7
33.50	1.02282	1528.3
34.00	1.02320	1528.9
34.50	1.02358	1529.4
34.69	1.02372	1529.7
35.00	1.02395	1530.0

*Notes:

1. Density: "Sigma-t" at 23°C and indicated salinity (LaFond, 1951, Table X).
2. Sound speed: entering tables for the speed of sound in seawater (NAVOCEANO SP-58 or SP-68) at 23°C, 1 atmosphere pressure (1.03 kg/cm²), and indicated salinity.
3. Salinity of 34.69 ppt is shown because it is a representative value for Central Pacific bottom water.

the Central Pacific can be used with negligible error over most of the deep Central Pacific. For those who cannot easily get such profiles for an area of interest, they can be constructed by entering tables for the speed of sound in seawater (NAVOCEANO, 1962, or 1966) using temperature, pressure, and salinity information from an atlas, report, or textbook (e.g., Defant, 1961, v. 1). Figure 2, for the Central Pacific, was constructed from the last source.

In determining the bottom-water velocity from a velocity-vs.-depth profile for the water column, a true depth, and not an echo-sounder depth, is needed. Most echo-sounders in United States oceanographic and Navy vessels are set at a water speed of 4800 ft/sec (1463 m/sec); the depths and contours printed on almost all official charts (Navy and C&GS) are echo-sounder depths, which are shallower than the true depths and must be corrected accordingly.

Two methods are currently used to correct echo-sounder to true depths: (1) by using a correction diagram for a particular area which is derived from Nansen cast data, or (2) by making the correction from Mathews' Tables (NAVOCEANO SP-68, table 11). Although Mathews' tables were published in 1939, they are surprisingly accurate in comparison with the large amounts of recent data. An example of a diagram, with depths in meters, which can be used in the first method is given in figure 3 (from Belshe, 1967) for an area off Hawaii. Figure 4, showing depths in fathoms, is a similar diagram for the same general area, based on a Nansen cast taken from R/V CARNEGIE in the 1930's. Although both figures are close, Belshe's diagram should be more accurate because it was based on more data. Either diagram can be used with negligible errors over most of the Central Pacific. Because of local interest, figure 5, showing depth corrections for the San Diego Trough, is included, as is a table (table A-6) listing various properties of the water column vs. depth for the same area. The diagram and table are accurate below the depths of seasonal changes in the upper waters.

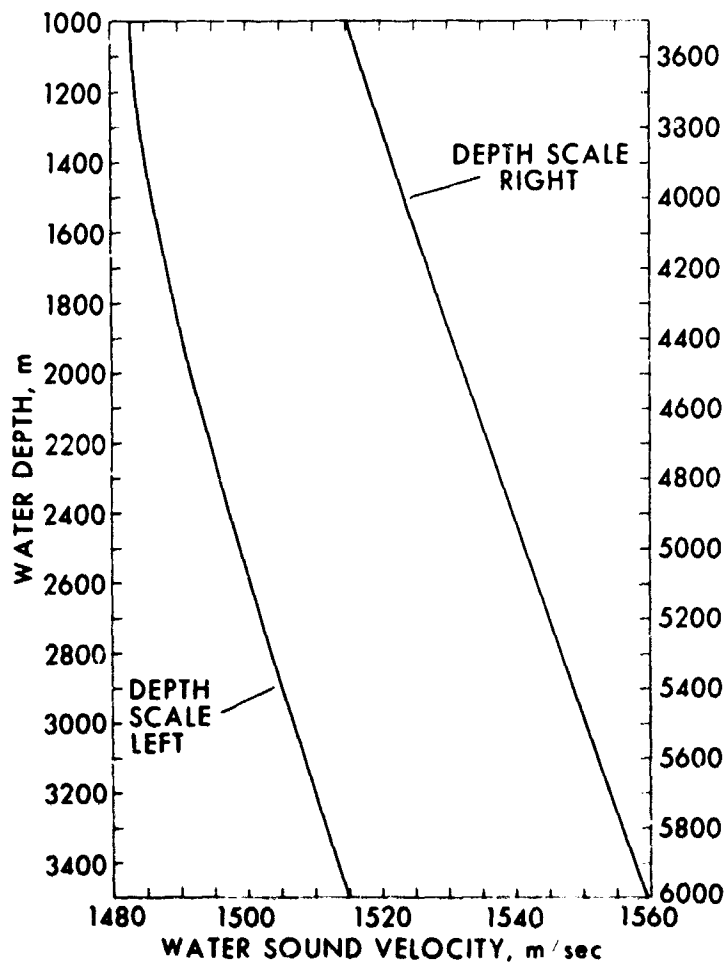


Figure 2. Sound velocity vs. water depth, Central Pacific.

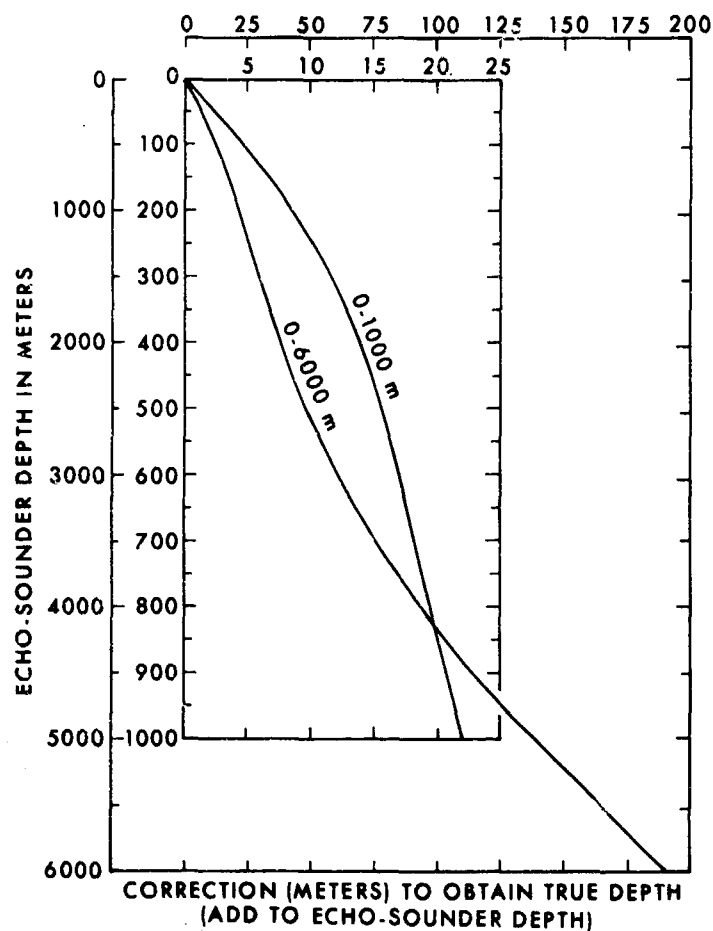


Figure 3. Correction to echo-sounder depth (m) to obtain true depth, Hawaiian area. (Belshé, 1967)

The following outline summarizes, step by step, the procedure discussed above, and in preceding sections, to determine the *in situ* sound velocity at the sediment surface (V_0); computations for specific examples are given for both the "ratio method" and "laboratory-correction method."

- A. To get sediment-surface sound velocity, *in situ* (V_0); assuming no information concerning sediment properties (using the ratio method).
 1. Determine sedimentary environment or province; e.g., abyssal hill (pelagic), continental terrace (shelf and slope).
Example: Location places the area in the "Abyssal Hill (Pelagic)" environment.
 2. Predict the sediment type within the environment.
Example: Table A-2 lists sediment properties for the abyssal hill (pelagic) environment; select the most common sediment type: silty clay (deep-sea "red" clay).

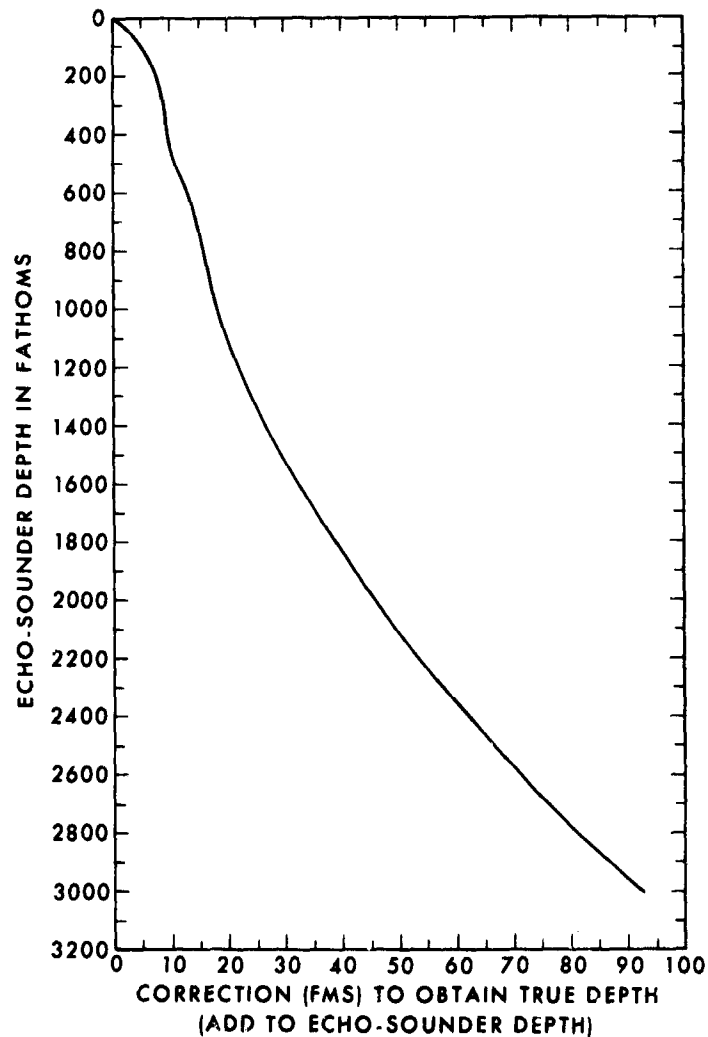


Figure 4. Correction to echo-sounder depth (fathoms) to obtain true depth, Hawaiian area.

3. Determine the average ratio (V_p sediment/ V_p seawater) for the sediment type within the environment.

Example: Table A-2 lists the average ratio for silty clay in the abyssal hill (pelagic) environment as 0.985.

4. Determine the true water depth in meters.

Example: the echo-sounder depth is 2658 fathoms; to correct to true depth:

- a. use NAVOCEANO SP-68, from Matthews (1939); the area is near Hawaii, which places it in Area 42 (p. 67).
- b. Table 11b, p. 93, Area 42: for an echo-sounder depth of 2658 fathoms, the correction is 76 fathoms (interpolated between 2600 and 2700 in

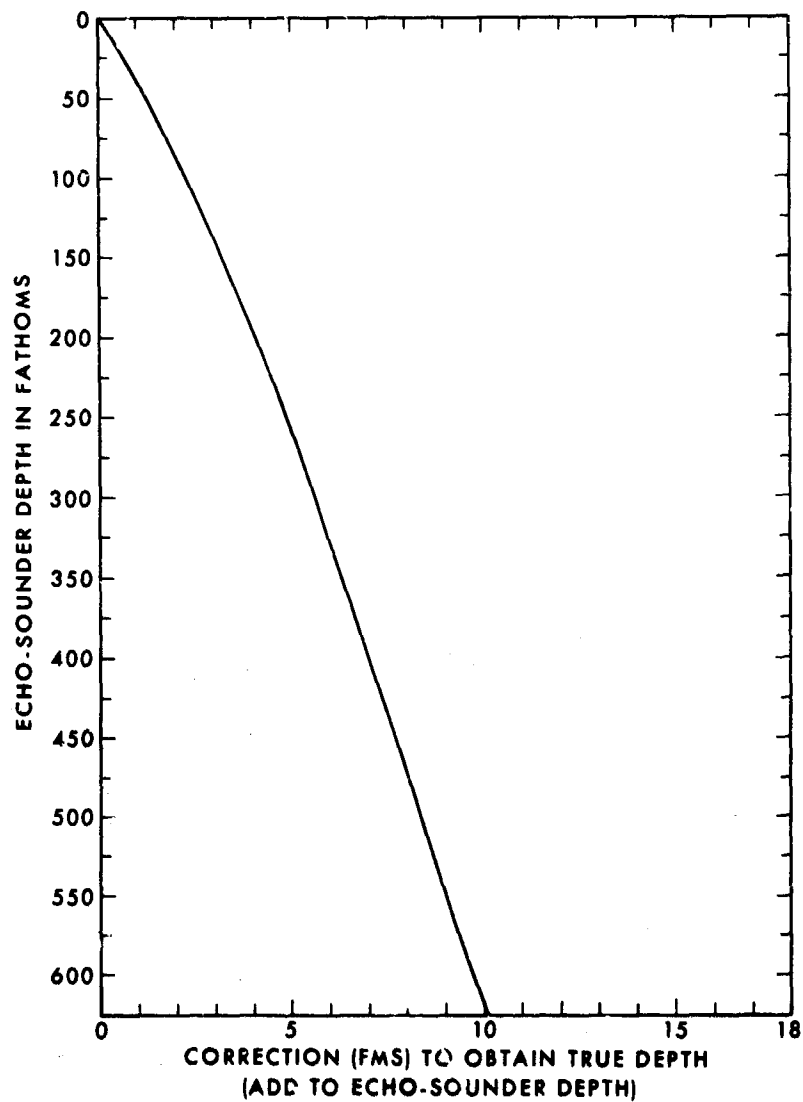


Figure 5. Correction to echo-sounder depth (fathoms) to obtain true depth, San Diego Trough.

the column headed "800," which means an echo-sounder set for 800 fathoms/sec, or 4800 feet/sec.

c. true depth is 2658 plus 76: 2734 fathoms.

d. convert to true depth in meters (from a table or by the conversion: 1 fathom = 1.8288 m); $2734(1.8288) = 5000$ m.

Note: For this particular area a slightly more accurate correction could be obtained from Belshé (1967) which was based on recent multiple Nansen casts; figure 3 indicates the correction to be about 132 m,

rather than 139 m, obtained from NAVOCEANO SP-68; the example used SP-68 because more accurate correction curves are not usually available for a desired area.

5. Determine sound velocity in the bottom water at the true depth from:
 - a. Profile of velocity vs. depth-in-water based on nearest Nansen cast (may need to extrapolate curve to desired depth).
 - b. Profile of velocity vs. depth constructed from temperature, pressure, and salinity information for the general area from an atlas, report, or textbook, and tables for the speed of sound in seawater.
 - (1) *Example:* A velocity-vs.-depth profile for the area was constructed from temperature, pressure, and salinity data in Defant (1961, v. 1) and Barbee (1965), and sound-speed tables in NAVOCEANO SP-58 or 68 (fig. 2).
 - (2) At a true depth of 5000 m (fig. 2) the water velocity (or bottom-water velocity) is 1542 m/sec.
6. Determine the sediment velocity, *in situ*, by multiplying the ratio (3, above) by the bottom-water velocity (5, above).
 - a. *Example:* sediment-surface sound velocity, *in situ* (V_0) = $0.985(1542) = 1519$ m/sec.

Having determined the ratio and attained a velocity-vs.-depth profile for the water column, the investigator is able to compute a sediment-surface velocity anywhere in the general area (with the aid of a contoured chart of the bottom) by determining the bottom-water velocity at the indicated depth and applying the ratio (which will hold over the area if the sediment type is the same). The key, of course, is that on top of a hill or in a depression both the bottom water and the sediment change in velocity, but the ratio remains the same.

- B. To determine the sediment-surface sound velocity (same example as in A), *in situ* (V_0); assuming no information concerning sediment properties (correction-to-laboratory-velocity method).
 1. and 2. Determine the sedimentary environment, or province, and sediment type.

Example: Same as in A., 1 and 2, above: abyssal hill (pelagic); use the most common sediment type, silty clay.
 3. Determine the laboratory sound velocity at a specific temperature and 1 atmosphere pressure.

Example: Table A-2 lists 1507 m/sec for abyssal hill (pelagic) silty clay at 23°C and 1 atmosphere pressure.
 4. Determine the true water depth in meters.

Example: As above (A., 4.): 5000 m.
 5. Determine temperature and pressure, *in situ*, of the bottom water in the area of interest; use specific data if available; if not, from general tables or figures.

Example: at 5000 m in the Central Pacific

 - a. Temperature: 1.5°C. (Barbee, 1965): NAVOCEANO SP-68 (table 7, p. 60) has temperature vs. depth to 4000 m (from Defant, 1961).

b. Pressure:

- (1) NAVOCEANO SP-68, figure 4, p. 46: pressure at 5000 m = $5000(0.10395) = 519.8 \text{ kg/cm}^2$.
- (2) Wilson (1959) has published a set of tables of pressure vs. depth at various salinities and temperatures; those for salinities of 34 and 35 ppt are included in table A-5. For the deep Central Pacific, the nearest table is that for 35 ppt and 0°C (table A-5b). For the San Diego Trough, one would use table A-5a (for a salinity of 34 ppt and 10°C).

6. Correct laboratory sediment velocity to *in situ* velocity.

Example (using tables for the speed of sound in seawater: NAVOCEANO SP-58, or table 12 in SP-68):

- (1) Temperature correction (SP-58, table 5; SP-68, table 12E)

$$\Delta V_t \text{ at } 23^\circ\text{C} \text{ ----- } +80.7 \text{ m/sec}$$

$$\Delta V_t \text{ at } 1.5^\circ\text{C} \text{ ----- } + 6.8$$

Difference (to be subtracted): 73.9 m/sec

- (2) Pressure correction

$$\Delta V_p \text{ at 1 atmosphere (1.03 kg/cm}^2\text{) --- } 0.2 \text{ m/sec}$$

$$\Delta V_p \text{ at 519.8 kg/cm}^2 \text{ ----- } \underline{86.2}$$

Difference (to be added) ----- 86 m/sec

- (3) Computation of *in situ* sediment velocity

laboratory velocity, 23°C , 1 atmosphere --- 1507 m/sec

$$\Delta V_t \text{ ----- } -74$$

$$\Delta V_p \text{ ----- } \underline{+86}$$

in situ velocity ----- 1519 m/sec

- Notes:
1. ΔV_p at various pressures can be determined from NAVOCEANO SP-58 (table 1) or SP-68 (table 12A) by subtracting the velocity for $V_0 + V_p$ at 1.03 kg/cm^2 (1 atmosphere pressure) from the indicated value for the *in situ* pressure.
 2. Table 2, NAVOCEANO SP-58, or table 12B (SP-68), lists V_0 (water) corrected for changes in depth ($V_0 + V_p$) with pressures derived assuming 0°C and 35 ppt; for more precise values, table 1 (SP-58) or 12A (SP-68) can be entered with a known pressure, or one derived from Wilson's (1959) pressure tables.

Although it is not claimed that the data for the North Pacific sediment types have universal application, it is probable that the various averaged values will be close to similar sediment types in other oceans. For example, Houtz and Ewing (1964) reported wide-angle reflection measurements of *in situ* sediment-surface velocities at four stations in the western Atlantic (Stations 4, 5, 8, and 11). Three of these (Stas. 4, 5, and 8) were probably in pelagic silt-clay; the average measured velocity in the sediment surface was

1522 m/sec. The "ratio method" of predicting sediment sound velocity *in situ* was applied to these areas as follows:

1. Water depth was determined from Houtz and Ewing (1964; table IV).
2. Bottom-water velocity at the correct depths was determined from NAVOCEANO TR-171 (1965; Region II-2, Summer: fig. 9, p. 17).
3. The ratio (velocity sediment/velocity water) for Pacific abyssal-hill silty clay (0.985) was multiplied by the bottom-water velocity at each station; the resulting average, predicted sediment velocity for the three stations was 1529 m/sec, or 7 m/sec (about 0.5 percent) greater than measured by Houtz and Ewing (1522 m/sec).

Porosity

The porosity of sediments in the sea floor is very slightly greater than in the laboratory because of the effect of hydrostatic pressure in reducing the volume of mineral grains. This increase is so small that it can be disregarded; it amounts to less than 0.001 percent for minerals and porosities usually encountered at pressures around 1000 kg/cm² (depths about 9500 m). Although true porosities do not require corrections from laboratory to *in situ* values, most published laboratory values require corrections to obtain true porosities. The literature values requiring corrections include water contents (percent of dry weight) and bulk densities of mineral grains ("dry density") from which porosities can be computed. These corrections involve allowance for the weight of dried salts oven-evaporated from the seawater within sediment pore-spaces and weighed with the dried, mineral solids during laboratory sediment analyses. The salt correction to porosity is small, but should be made when laboratory procedures are precise, especially when porosity is to be used in computations of elastic constants.

In a saturated sediment, the volume of voids (or pore space) occupied by water is expressed as porosity, *n*, or void ratio, *e* (more common in the literature of soil mechanics):

$$n = \frac{\text{Volume of voids}}{\text{Total volume}} = \frac{e}{1 + e}$$

$$e = \frac{\text{Volume of voids}}{\text{Volume of solids}} = \frac{n}{1 - n}$$

Porosity is usually expressed in percent, and void ratio as a decimal fraction.

The causes of porosity in marine sediments and the interrelations of porosity with other properties were discussed in Part I. Average values are in tables A-1 and A-2.

Porosity is usually computed assuming that 1 gram of oven-evaporated water equals 1 cc (true for distilled water at room temperatures when density is rounded to the second decimal place). Thus, porosity, in the usual laboratory computations, is equal to the weight of evaporated water divided by the volume of saturated sediment; the resulting value is usually listed without correction for the dried salts weighed with the mineral residue. This correction amounts to an increment to porosity varying, according to porosity, from 0.5 to 1 percent. Because porosity has an important influence on the acoustic and elastic properties of marine sediments, the salt correction should be made; three methods are detailed below.

Given: (1) salinity of pore water (use salinity of bottom water), (2) weight of oven-evaporated water, (3) weight of dried minerals (with salt), (4) temperature at time of weighing wet sediment.

A. Volume-of-seawater method.

$$1. \text{ Weight of seawater} = \frac{\text{weight of evaporated water}}{1 - (\text{salinity, ppt})}$$

$$\text{Note: "(\text{salinity, ppt})" for water salinity of 35 ppt} = \frac{35}{1000} = 0.035$$

$$1 - (\text{salinity, ppt}) = 1.0 - 0.035 = 0.965.$$

$$2. \text{ Volume of seawater} = \frac{\text{wt. of seawater}}{\text{density of seawater, 1 atmos., lab temp.}}$$

$$3. \text{ Porosity} = \frac{\text{volume of seawater}}{\text{total volume of saturated sample}}$$

Note: in a unit volume (1 cc), porosity is equal to the volume of seawater.

B. Volume-of-solids method: given a measured or assumed value for bulk density of solids.

1. Determine weight of seawater (A., 1., above).

2. Weight of salt = wt. of seawater - wt. evaporated water.

3. True wt. of solids = wt. oven-dried minerals - wt. of salt.

$$4. \text{ Volume of solids} = \frac{\text{true wt. of solids}}{\text{density of solids'}}$$

or volume of solids = total volume - volume of seawater.

5. Volume of voids, or pore space = total volume - volume of solids.

$$6. \text{ Porosity} = \frac{\text{volume of voids}}{\text{total volume}}$$

Notes: Factors in the above computations which are not, or may not be, measured are density of seawater and the bulk density of minerals (solids). These can be derived as follows:

- Laboratory (1 atmosphere pressure) density of seawater (Sigma-t) varies with salinity and temperature. It can be determined from Sigma-t tables, including "Determining density of seawater" (NAVOCEANO SP-68, table 10, p. 302; not properly indexed). Values of seawater density (Sigma-t) at 23°C, 1 atmosphere pressure, and various salinities are listed in table 2.
- The bulk density of mineral solids ("dry density") must be (1) computed with a salt-free, dry, mineral residue, (2) measured with a pycnometer, or other, method in which salts are elutriated from the oven-dried materials, or (3) a value assumed. A value of 2.66 g/cc is usually assumed in soil-mechanics computations involving sediments and soils on land; average values for North Pacific sediments are given in table A-1.

C. Empirical corrections.

The amount of dried salt weighed with the dried minerals varies according to porosity, or the amount of seawater. As a result of numerous corrections and computations it was determined that a porosity with salt can be multiplied by 1.012 to get salt-free, or "true," porosity. This empirical correction can be applied with little or no significant difference from the results of more elaborate computations of A. and B., above. Table 3 was derived by multiplying a given porosity (with salt) by 1.012 to obtain salt-free porosity.

TABLE 3. EMPIRICAL CORRECTIONS TO SEDIMENT POROSITY
TO OBTAIN SALT-FREE POROSITY

Empirical Salt Correction ¹		Pore-Water Salinity Known	
Porosity With Salt (%)	Porosity Increment (%)	Salinity of Pore Water (ppt)	Multiplier ²
35.0 - 37.4	0.4	33.00	1.01143
37.5 - 45.8	0.5	33.50	1.01158
45.9 - 54.1	0.6	34.00	1.01173
54.2 - 62.4	0.7	34.50	1.01187
62.5 - 70.8	0.8	34.69	1.01195
70.9 - 79.1	0.9	35.00	1.01203
79.2 - 87.4	1.0		
87.5 - 96.0	1.1		

¹ Correction to allow for salts oven-evaporated from pore water; derived by multiplying porosity, with salt, by 1.012.

² Multiply sediment porosity, with salt, by "multiplier" to get salt-free porosity.

Density

There is a small, and probably insignificant, correction of laboratory values of sediment saturated bulk density to *in situ* values. This correction involves an increment to density resulting from more dense water in the pore spaces of the sediment in the sea floor. Published laboratory values of saturated bulk density ("wet density") can usually be used as *in situ* values without correction because the correction may be within the margins of error in laboratory procedures, or insignificant in view of the lateral and vertical changes in the sea floor over small areas and depths. However, the correction is simple and can easily be applied by inspection or computations.

Laboratory values of saturated bulk densities, bulk densities of mineral grains, porosities, and interrelationships between these properties were discussed in Part I. Tables A-1 and A-2 list the average laboratory values of these properties for the sediment types within each North Pacific environment.

The saturated bulk density (hereinafter referred to as "density") of a unit volume of gas-free sediment has two components: mineral grains and water within pore spaces. The relationships between these constituents and porosity are

$$\rho_{mt} = .n \rho_w + (1 - .n) \rho_s \quad (1a)$$

or

$$.n = \frac{\rho_s - \rho_{mt}}{\rho_s - \rho_w} \quad (1b)$$

$$\rho_s = \frac{\rho_{mt} - .n \rho_w}{(1 - .n)} \quad (1c)$$

$$\rho_w = \rho_s - \left(\frac{\rho_s - \rho_{sat}}{.n} \right) \quad (1d)$$

where

ρ_{sat} = saturated bulk density (or "density")

$.n$ = porosity (decimal fraction); volume of voids/total sample volume

ρ_w = density of water in void spaces (or "pore spaces")

ρ_s = bulk density of mineral solids (or "grain density")

Given salt-free porosity and bulk density of minerals, the only variable factor between laboratory and *in situ* density is density of the pore water. Table 2 lists laboratory values for the density of seawater at 23°C and various salinities; table 1 lists *in situ* values typical in the Central Pacific at various water depths. At a given salinity the differing values are due to the effects of temperature and pressure. The near-maximum effect of these values on density (at usual water depths) can be illustrated by the following example.

Given: a silty clay from the abyssal hill (pelagic) environment with a bulk grain density, ρ_s , of 2.70 g/cc, a porosity, $.n$, of 80 percent, and a pore-water salinity of 34.69 ppt.

- a. In the laboratory, water density, ρ_w , is 1.0237 g/cc (table 2); from equation (1a),

$$\rho_{sat} = .8(1.0237) + (.2)2.70 = 1.359 \text{ g/cc}$$

- b. In the sea floor at a water depth of 6000 m, $\rho_w = 1.0550$ g/cc (table 1),

$$\rho_{sat} = .8(1.055) + .2(2.70) = 1.384 \text{ g/cc}$$

The difference between laboratory and *in situ* densities is 0.025 g/cc; a simpler computation of the difference is: $.n\rho_w$, *in situ*, minus $.n\rho_w$, lab.

Most bottom-water salinities are between 34.50 and 35.00 ppt in the open oceans; therefore, a laboratory water density of 1.024 g/cc (table 2) can be used without significant error in laboratory computations or in determining the difference between laboratory and *in situ* values.

The maximum increments to density are in deep-water, high-porosity sediments, which are usually between 75 and 85 percent. Consequently, if sediment densities are rounded to the second decimal place, table 4 can be used to correct, by inspection, laboratory densities to *in situ* values. For example, the average laboratory density and porosity of abyssal-hill silty clay is 1.37 g/cc and 79.4 percent; an approximate average depth in the deep Central Pacific is 5000 m; thus, by inspection (table 4), the increment to density is 0.02 g/cc, and the average *in situ* density is 1.39 g/cc.

Given only one value, either density or porosity, the missing value can be derived in several ways: enter a density-vs.-porosity diagram, use a regression equation for such diagrams, or compute a value (if grain density is known). The last method is most accurate. For example, given abyssal-hill silty clay with a grain density of 2.71 g/cc, and a porosity of 79.4 percent, and using equation 1a and a laboratory water density of 1.024 g/cc, a computed value for laboratory density of 1.37 g/cc results. This type of computation, and those indicated by the variations of equation 1a, i.e., 1b-1d, can be used to verify the accuracy of published data, or to cross-check the accuracy of independent laboratory measurements of density, porosity, and grain density (as in the above example from table A-2).

TABLE 4. EMPIRICAL CORRECTIONS (TO BE ADDED)
TO LABORATORY DENSITY OF HIGH-POROSITY SEDIMENTS
TO OBTAIN *IN SITU* VALUES.

Water Depth (m)	Density Increment (g/cc)
<u>Porosity \approx 75%</u>	
0 - 500	0.00
600 - 3400	0.01
3500 - 6500	0.02
6600 - 9900	0.03
> 9900	0.04
<u>Porosity \approx 85%</u>	
0 - 300	0.00
400 - 2900	0.01
3000 - 5600	0.02
5700 - 9400	0.03
> 9400	0.04

Impedance

The characteristic impedance of a medium is the product of density, ρ , and compressional-wave velocity, V_p (impedance = ρV_p , g/cm² sec). This important property determines the amount of energy reflected when sound passes from one medium into another. An echo sounder or reflection profiler merely records time vs. impedance mismatches.

Impedances are used in computations of reflection coefficients and bottom losses of sound energy incident on the sea floor. For the sediments of this report, impedances were computed at laboratory values (23°C, 1 atmosphere) and discussed and tabulated in Part I; these tables are included in Appendix A (tables A-1 and A-2).

Laboratory impedances require correction to *in situ* values. No discussion is required in this section because methods of correcting both velocity and density to *in situ* values were discussed in previous sections; the *in situ* impedance is merely the product of the corrected values.

Impedance of seawater, *in situ*, at various depths in the Central Pacific, and in the laboratory, can be computed from the density and velocity values in tables 1 and 2.

An example of the difference between laboratory and *in situ* values of sediment impedance is noted below.

Given: an abyssal-hill silty clay from 5000-m water depth.

1. Sediment density and velocity values from table A-2; laboratory impedance, density (velocity)

$$= 1.37 \text{ g/cc } (1.507 \text{ cm/sec} \times 10^5) = 2.0646 \text{ g/cm}^2 \text{ sec} \times 10^5.$$

2. The corrections of density and velocity from laboratory to *in situ* conditions for this example were noted in previous sections; *in situ* impedance = 1.39 g/cc (1.519 cm/sec $\times 10^5$) = 2.1114 g/cm² sec $\times 10^5$.

Reflection Coefficient and Bottom Loss at Normal Incidence

The Rayleigh reflection coefficient for a simple, harmonic, plane wave incident on a plane boundary between two fluids across which there is a density and velocity change expresses the ratio of amplitudes, or pressures, of a reflected wave to that of the incident wave (references and discussion in Part I); at normal incidence, the Rayleigh reflection coefficient, R , is expressed by

$$R = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1} \quad (2)$$

where

$\rho_1 V_1$ = impedance of the first medium (water, in the present case)

$\rho_2 V_2$ = impedance of the second medium (sediment, in the present case)

Bottom loss, BL , of a plane wave at normal incidence (on a peak-pressure basis), expressed in dB, is

$$BL = 20 \log R \quad (3)$$

In Part I, Rayleigh reflection coefficients and bottom losses at normal incidence were computed for laboratory conditions of 23°C and 1 atmosphere; the results were listed in tables (reproduced as tables A-1 and A-2) and relationships with other properties shown in scatter diagrams.

As discussed in Part I, and by Hamilton *et al.* (1969), the author favors, for the sea floor, a viscoelastic model in which rigidity is present, and believes that the fluid/fluid model involved in the preceding equations is less useful to explain all aspects of bottom loss. The Rayleigh reflection coefficients and bottom losses at normal incidence were listed and discussed because they appear to be close to values measured at sea, and for possible usefulness in studies of the reflection and refraction of sound.

Computations of bottom losses for laboratory and *in situ* conditions indicate that the laboratory values shown in tables A-1 and A-2 are within a few tenths of a dB of values computed for *in situ* conditions. These computations involve only sediment and water impedances (equations 2 and 3) and their corrections to *in situ* values which were discussed in a previous section; therefore, the corrections of reflection coefficients and bottom losses need not be further elaborated.

Although reflection coefficients and bottom losses at normal incidence for *in situ* conditions can be easily computed, table 5 indicates that the laboratory values are so close to *in situ* values that, usually, the laboratory averages listed in tables A-1 and A-2 can be used as average *in situ* values for generalized studies. As discussed and illustrated in Part I, these averages are close to actual measurements at sea by Breslau (1967) and Fry and Raitt (1961).

TABLE 5. COMPARISON BETWEEN LABORATORY AND *IN SITU* VALUES OF RAYLEIGH REFLECTION COEFFICIENTS AND BOTTOM LOSSES AT NORMAL INCIDENCE.*

Environment and Sediment Type	Reflection Coefficient		Bottom Loss (dB)	
	Lab.	<i>In Situ</i>	Lab.	<i>In Situ</i>
<u>Continental Terrace (Shelf and Slope)</u>				
Fine sand	0.3763	0.3751	8.5	8.5
Clayey silt	0.1674	0.1633	15.5	15.9
<u>Abyssal Plain (Turbidite)</u>				
Silty clay	0.0927	0.0876	20.7	21.5
<u>Abyssal Hill (Pelagic)</u>				
Silty clay	0.1373	0.1316	17.2	17.6

*Notes:

1. Laboratory values for sediment properties from tables A-3 and A-4; laboratory, water impedances computed from table 2.
2. *In situ* conditions assumed to be at 11 m (fine sand) and 1100 m (clayey silt) off San Diego (see table of variations of water properties with depth off San Diego, Appendix, table A-6), and at 5000-m water depth for deep-water silty clays (table 1).

Elastic Constants

Part II (Elasticity and Elastic Constants) included discussions of elastic and visco-elastic models for water-saturated porous media, and measurements and computations of elastic constants including density, compressional-wave velocity, compressibility, incompressibility (bulk modulus), rigidity (shear) modulus, Lamé's constant, Poisson's ratio, and shear-wave velocity in marine sediments. One of the conclusions of Part II is that the equations of elasticity can be used to compute unmeasured elastic constants in the case of water-saturated sediments.

To compute elastic constants with the equations of elasticity, density and any two other elastic constants are required. In the case of marine sediments, the density and compressional-wave velocity can be easily measured or predicted; the best third constant would be the velocity of shear waves, at least for purposes of underwater sound and geophysics.

Information on the velocity of shear waves in marine sediments is rare. Those known to the writer are assembled and discussed in Part II. Lacking sufficient information on shear-wave velocities, the third constant selected for use in computing the other constants was the bulk modulus (or incompressibility) of the sediment, water-mineral system. This constant was selected because it appears possible to compute, in a logical manner, the bulk modulus of the sediment from its components, without estimations. The theoretical basis of this computation follows that of Gassmann (1951); see Part II for an extended discussion.

The components of the computed system bulk modulus κ , are porosity, n , the bulk modulus of pore water, κ_w , an aggregate bulk modulus of mineral grains, κ_g , and a

bulk modulus of the sediment structure (or frame), κ_f , formed by the mineral grains. Good values for the bulk modulus of distilled and seawater, and most of the common minerals of sediments, have been established in recent years. This leaves only a value for the frame-bulk modulus, κ_f , needed to compute a bulk modulus for the water-mineral system (following Gassmann, 1951).

A contribution of Part II was the derivation of a relationship between sediment porosity and the dynamic frame-bulk modulus. Using this relationship, the frame-bulk modulus was derived for each sample and used with the bulk moduli of pore water and minerals to compute the system bulk modulus. This computed bulk modulus, and measured density and compressional-wave velocity were then used to compute the other elastic constants. These measured and computed values were listed in tables according to sediment type and environment (Part II, tables 1 and 2; included in the Appendix of this report as tables A-3 and A-4). All values were, as in Part I, referred to 23°C and 1 atmosphere pressure.

The values for elastic constants at 23°C and 1 atmosphere in the tables and scatter diagrams of Part II are usable for basic studies and interrelationships, but cannot be used as *in situ* values because all three of the constants (ρ , V_p , and κ) used in the computations require correction from laboratory to *in situ* conditions.

The corrections to laboratory values of density and compressional-wave velocity to get *in situ* values have been discussed in preceding sections. The components used in computing the system bulk modulus, κ (except porosity) also require corrections. The major change is in the bulk modulus of pore water, κ_w , because of the effects of temperature and pressure on the large volume of pore water relative to the volume of minerals in deep-water, high-porosity sediments.

The aggregate bulk modulus of mineral grains, κ_s , increases with higher hydrostatic pressure and lower temperature, but the effect of this change is so small (and masked by the larger change in κ_w) in the range of temperatures and pressures involved, that it can be disregarded, and laboratory values of κ_s can be used as *in situ* values. The magnitude of these small changes in κ_s due to temperature and pressure changes from laboratory to *in situ* conditions can be derived from studies on quartz (Soga, 1968) and calcite (Peselnick, 1962; Peselnick and Wilson, 1968).

The bulk modulus of the sediment mineral frame, κ_f , depends on porosity or volume of minerals and intergranular pressure (buoyed weight of mineral grains; see discussion in Part II). Although the buoyed weight of mineral grains is very slightly less *in situ* because pore water is more dense *in situ*, the change in intergranular pressure is insignificant, and the same value for κ_f can be used in the laboratory and *in situ*.

In summary, prior to computing *in situ* elastic constants, the following properties require correction from laboratory to *in situ* conditions: density, compressional-wave velocity, and the bulk modulus of pore water. Porosity, the aggregate bulk modulus of minerals, and the frame-bulk modulus require no change.

Although not essential, study of Parts I and II of this report is recommended prior to computations of *in situ* elastic constants for sediments. Elaboration of the many factors involved in such computations would cause considerable duplication and unduly lengthen this Part.

The laboratory values of density, porosity, and compressional-wave velocity are critical in computations of elastic properties as outlined in this section. When laboratory measurements are carefully made, the values for density and salt-free porosity should be cross-checked with grain density by using equation 1a. Usually, the values of density and fractional porosity should be carried to three decimal places. Sediment compressional-wave velocities should be rounded to the nearest m/sec. Water densities and sound speeds are accurately determined from oceanographic tables and measurements and can be used as tabulated.

As a practical consideration in the computations, it is recommended that the cgs system be used and that all values be reported $\times 10^{10}$; thus most of the elastic constants are reported in dynes/cm² $\times 10^{10}$. This is convenient because compressional velocities are usually reported in m/sec, which, when converted to cm/sec by inspection, result in $1.\text{xxx} \times 10^5$, and ρV_p^2 , an important value in elasticity, results in dynes/cm² $\times 10^{10}$ when g/cm sec² is converted to dynes/cm². In computing velocity of shear waves, the square root of (μ/ρ) is taken, so that if $\times 10^{10}$ is used, the result, $(\times 10^{10})^{1/2}$, is cm/sec $\times 10^5$, and the decimal can be moved by inspection to get m/sec.

All the laboratory measurements (density, porosity, and compressional-wave velocity), and components of the system bulk modulus (moduli of pore water, minerals, and frame) used in computations of the elastic constants have margins of error; some are known (see Part I), and some are unknown. Consequently, no attempt was made to statistically estimate variances or errors in the final computations of elastic constants. The numbers of decimal places shown in the examples and table are for purposes of comparison between the various computations and sediment types, and should not be taken as the author's estimates of accuracy. The values of the elastic constants computed and listed should be considered as approximations and predictions of these values when more, future, measurements (such as for the velocities of shear waves) are available.

OUTLINE OF PROCEDURES TO COMPUTE *IN SITU* VALUES OF ELASTIC CONSTANTS FROM CORRECTED, LABORATORY, SEDIMENT DATA

Examples of the computations are given for three sediment types — two from the continental terrace, and one from the deep-sea abyssal-hill environment.

Given:

1. Fine sand from the continental shelf off San Diego from a water depth of 11 m.

Sediment properties, laboratory:

- a. Ratio, velocity in sediment/velocity in seawater = 1.14
- b. Porosity, n = 44.0 percent
- c. Grain density, ρ_s = 2.70 g/cc.

2. Clayey silt: a turbidite from the Continental Borderland (San Diego Trough) at a water depth of 1100 m.

Sediment properties, laboratory:

- a. Ratio = 1.000
- b. Porosity, n = 75.0 percent
- c. Grain density, ρ_s = 2.71 g/cc.

3. Silty clay ("red clay") from the Central Pacific, abyssal-hill environment at a water depth of 5000 m.

Sediment properties, laboratory:

- a. Ratio = 0.985
- b. Porosity, n = 79.4 percent
- c. Grain density, ρ_s = 2.71 g/cc.

Note: Values of water properties, *in situ*, used below in connection with fine sand and clayey silt (examples 1 and 2) were derived from Nansen-cast data off San Diego (Appendix, table A-6) at 11 and 1100 m; for silty clay (example 3) from table 1 at 5000 m.

1. Determine sediment compressional-wave (sound) velocity *in situ*, using procedures discussed in the "Sound Velocity" section.

Examples:

- a. Fine sand at 11 m.

- (1) Bottom-water sound velocity = 1505.65 m/sec at 11 m water depth.
- (2) *In situ* sediment velocity = ratio \times (bottom-water velocity) = $1.14(1505.65) = 1716$ m/sec.

- b. Clayey silt at 1100 m.

- (1) Bottom-water velocity = 1483.64 m/sec.
- (2) *In situ* sediment velocity = $1.000(1483.64) = 1484$ m/sec.

- c. Silty clay at 5000 m.

- (1) Bottom-water velocity = 1541.7 m/sec.
- (2) *In situ* sediment velocity = $0.985(1541.7) = 1519$ m/sec.

2. Determine *in situ* sediment density using procedures discussed in "Density" section.

Examples (using procedures when porosity, n , and grain density, ρ_s , are known):

- a. Fine sand at 11 m.

- (1) Density of bottom water = 1.02490 g/cc.
- (2) *In situ* sediment density, $\rho_{sat} = n\rho_w + (1 - n)\rho_s = 0.44(1.0249) + (0.56)2.70 = 1.963$ g/cc.

- b. Clayey silt at 1100 m.

- (1) Density of bottom water = 1.03257 g/cc.
- (2) *In situ* sediment density = $0.75(1.03257) + (0.25)2.71 = 1.452$ g/cc.

- c. Silty clay at 5000 m.

- (1) Density of bottom water = 1.05069 g/cc.
- (2) *In situ* sediment density = $0.794(1.05069) + (0.206)2.71 = 1.393$ g/cc.

3. Determine *in situ* values of the component-bulk moduli (pore water, κ_w , aggregate of mineral grains, κ_s , and the mineral frame, κ_f) necessary to compute an *in situ* sediment-bulk modulus, κ .

Examples:

- a. Bulk modulus of pore water, κ_w ; using the equation for sound velocity in a liquid: $V_p = (\kappa/\rho)^{1/2}$, or $\kappa = \rho V_p^2$, g/cm sec², or dynes/cm².

- (1) Fine-sand pore water, $\kappa_w = \rho V_p^2 = 1.0249$ g/cc $(1.50565 \text{ cm/sec} \times 10^5)^2 = 2.32343$ g/cm sec² $\times 10^{10}$, or dynes/cm² $\times 10^{10}$.
- (2) Clayey-silt pore water, $\kappa_w = 1.03257$ g/cc $(1.48364 \text{ cm/sec} \times 10^5)^2 = 2.27289$ dynes/cm² $\times 10^{10}$.
- (3) Silty-clay pore water, $\kappa_w = 1.05069$ g/cc $(1.5417 \text{ cm/sec} \times 10^5)^2 = 2.49732$ dynes/cm² $\times 10^{10}$.

b. Aggregate bulk modulus of mineral grains, κ_s . This bulk modulus was determined by the Voigt-Reuss-Hill averaging method (Hill, 1952, 1963), which involves the volumetric contributions of the bulk modulus of individual mineral species to the total bulk modulus of the aggregate of minerals (see Part II for discussion). The range of aggregate bulk moduli for most sediments is so small (see examples in Part II, Appendix B) that estimates can be used when mineralogy is not exactly known. For the three examples in this outline, aggregate bulk moduli for the fine sand and clayey silt off San Diego were determined from the mineralogy of these sediments in Emery *et al.* (1952) and Shepard and Einsele (1962); the bulk moduli of mineral species were derived from the literature (see Part II, Appendix B, for references). The aggregate-bulk modulus for deep-sea clay (example 3: $\kappa_s = 50$ dynes/cm² $\times 10^{10}$) was taken from Skempton (1961), who used this value for "clay".

For the three examples:

- (1) Fine-sand aggregate-bulk modulus of mineral grains, $\kappa_s = 52.33$ dynes/cm² $\times 10^{10}$.
 - (2) Clayey-silt aggregate-bulk modulus of mineral grains, $\kappa_s = 54.42$ dynes/cm² $\times 10^{10}$.
 - (3) Silty-clay aggregate-bulk modulus of mineral grains, $\kappa_s = 50$ dynes/cm² $\times 10^{10}$.
- c. Dynamic frame-bulk modulus, κ_f . One contribution of Part II was in derivation of a relationship between dynamic frame-bulk moduli and porosities (Part II, fig. 2, and regression equations for the data: Part II, Appendix C). These equations are (n = fractional porosity):

for sands: κ_f , dynes/cm² $\times 10^9$

$$\log \kappa_f = 2.71405 - 4.12135(n)$$

for silt-clays: κ_f , dynes/cm² $\times 10^8$

$$\log \kappa_f = 3.73807 - 4.25571(n)$$

These equations were used to derive the following dynamic frame-bulk moduli for the three examples:

1. Fine sand: $\kappa_f = 0.68802$ dynes/cm² $\times 10^{10}$
2. Clayey silt: $\kappa_f = 0.03530$ dynes/cm² $\times 10^{10}$
3. Silty clay: $\kappa_f = 0.02295$ dynes/cm² $\times 10^{10}$
4. Compute an *in situ* value for the water-mineral system-bulk modulus, κ , using Gassmann's equations (1951; discussed in Part II):

$$\kappa = \kappa_s \frac{\kappa_f + Q}{\kappa_s + Q} \quad Q = \frac{\kappa_w(\kappa_s - \kappa_f)}{n(\kappa_s - \kappa_w)} \quad (4)$$

where

κ_s = aggregate-bulk modulus of mineral solids
 κ_f = frame-bulk modulus ("skeletal" bulk modulus of Gassmann, 1951)

κ_w = bulk modulus of pore water

n = decimal-fractional porosity of sediment.

The above equation was used to compute a system-bulk modulus, κ , for each of the three examples, using the given measured porosities, and the components (κ_w , κ_s , and κ_f) derived above (item 3).

Examples:

a. Fine-sand system-bulk modulus, $\kappa = 5.5617 \text{ dynes/cm}^2 \times 10^{10}$

b. Clayey-silt system-bulk modulus, $\kappa = 3.0204 \text{ dynes/cm}^2 \times 10^{10}$

c. Silty-clay system-bulk modulus, $\kappa = 3.1252 \text{ dynes/cm}^2 \times 10^{10}$.

5. Use the equations of elasticity to compute *in situ* elastic constants using the measured, laboratory values of saturated bulk density, ρ_{sat} , and compressional-wave velocity, V_p , corrected to *in situ* values (1. and 2., above), and the computed *in situ* system-bulk moduli, κ (4., above). Those equations using these three constants are favored:

$$\text{Compressibility, } \beta = \frac{1}{\kappa} \quad (5)$$

$$\text{Lamé's constant, } \lambda = \frac{3\kappa - \rho V_p^2}{2} \quad (6)$$

$$\text{Poisson's ratio, } \sigma = \frac{3\kappa - \rho V_p^2}{3\kappa + \rho V_p^2} \quad (7)$$

$$\text{Rigidity (shear) modulus, } \mu = (\rho V_p^2 - \kappa)^{3/4} \quad (8)$$

$$\text{Velocity of shear wave, } V_s = (\mu/\rho)^{1/2} \quad (9)$$

$$\text{Young's modulus, } E = \frac{9\kappa(\rho V_p^2 - \kappa)}{\rho V_p^2 + 3\kappa} \quad (10)$$

These and other equations of elasticity, including variations, can be found in convenient tables by Gassmann (1951, p. 7) and Birch (1961, p. 2206; also in Birch, 1966, p. 100).

The *in situ* values of density, porosity, compressional-wave velocity, and the computed elastic constants (using the above equations) for the three examples are listed in table 6.

TABLE 6. *IN SITU* VALUES OF ρ , n , AND V_p , AND COMPUTED ELASTIC CONSTANTS*

Environment and Sediment Type	Laboratory Corrected ¹			Computed Elastic Constants ²⁻⁷					
	ρ	n	V_p	κ	β	λ	σ	μ	V_s
<u>Continental Terrace</u>									
Fine sand	1.963	44.0	1716	5.5617	0.1798	5.4523	0.485	0.1640	289
Clayey silt	1.452	75.0	1484	3.0204	0.3311	2.9318	0.478	0.1329	303
<u>Abyssal Hill</u>									
Silty clay	1.393	79.4	1519	3.1252	0.3200	3.0807	0.489	0.0667	219

*For examples in text, pp. 31.

Notes:

1. Density, ρ , g/cc; porosity, n , %; compressional-wave velocity, V_p , m/sec
2. Bulk modulus (incompressibility), κ , dynes/cm² $\times 10^{10}$
3. Compressibility, β , cm²/dynes $\times 10^{-10}$
4. Lamé's constant, λ , dynes/cm² $\times 10^{10}$
5. Poisson's ratio, σ
6. Rigidity (shear) modulus, μ , dynes/cm² $\times 10^{10}$
7. Shear-wave velocity, m/sec

PROPERTIES OF LAYERS

Introduction

This section will be concerned with the properties of relatively thin layers within soft, unlithified sediment bodies and those layers which may be expected to lie immediately under the unlithified sediments. Whether or not these layers reflect sound energy depends on the frequency of the sound wave and on the properties of the layers. Sound energy is reflected when there is sufficient impedance mismatch between adjacent layers, and the lower layer has the greater impedance. Characteristic acoustic impedance is the product of density and compressional velocity; therefore, any factor which increases either density or velocity may cause impedance mismatches and reflection of sound energy.

Although the exact location and properties of layers within sediment bodies cannot be predicted, it can be predicted that, in some areas, these layers will be present. It is useful for several purposes in the fields of underwater acoustics and geophysics to define the usual ranges of the properties of these layers.

When mineral particles from terrigenous sources, calcareous or siliceous particles from biogenic sources, or minerals formed in place on the sea floor, are cemented together they form mudstone, shale, limestone, and various siliceous rocks. The most usual cementing agents are calcium carbonate, silica, clay minerals, iron, manganese, and zeolites. These cementing agents are derived from mineral solution and redeposition, or precipitation from seawater. The cementation may be very slight, such as that forming mudstone in the walls of the La Jolla Submarine Canyon, and the indurated clays of the deep Pacific (Morgenstein, 1967), or very extensive, to form low-porosity limestones and cherts.

In the following sections, the various common layers and their properties will be briefly discussed with tables which detail the expectable ranges of these properties as terrigenous minerals are cemented to become mudstones and shales, siliceous sediments become diatomites, cherts and other siliceous rocks, and as calcareous sediments become limestones and dolomites. All of these rock types are known to underlie, in various areas, the soft, unlithified sediments of the sea floor. A special section will be devoted to a discussion of the properties of basalt, which is a common underlying rock type in all oceans. With few exceptions, all examples are from the sea floor or oceanic islands, seamounts, and ridges.

Tables 7, 8, 9, and 10 list representative values of density, porosity, and velocity for a wide variety of sediments and sedimentary rocks. These values are from cited literature and the writer's previously unpublished measurements (cited as: "Hamilton, this report"). All the sediment velocities are for 23°C (or "room temperature") and 1 atmosphere pressure; the corrections to obtain *in situ* velocities were discussed in a previous section. The effects of temperature and pressure on various rocks are reasonably well known, but will not be discussed in this section for reasons noted in the section on velocity gradients within the sea floor.

TABLE 7. REPRESENTATIVE EXAMPLES OF LAYERS IN MARINE SEDIMENTS FORMED BY TURBIDITES, MUDSTONES, AND SHALE.

Sediment Type	Density (g/cc)	Porosity (%)	Velocity (m/sec)	Area	Ref.*
<u>Sand (cores)</u>					
	2.09	35.2	1828	Newfound. Basin	1
	2.14	35.1	1832	Mediterranean	2
	1.96	43.1	1635	NW Atlantic	1
	1.86	49.7	1694	San Diego Trough	3
<u>Silty sand (cores)</u>					
	1.84	50.4	1700	La Jolla Fan	3
	1.92	47.4	1765	Asian Shelf	3
	1.84	51.3	1674	Japan Basin	3
	1.65	62.0	1631	Mediterranean	2
<u>Sandy silt (cores)</u>					
	2.01	38.8	1861	Japan Basin	3
	1.65	56.6	1622	Japan Basin	3
	1.67	61.3	1605	Bering Shelf	3
	1.81	52.8	1607	Aleutian Basin	3
	1.78	57.7	1611	Mediterranean	2
<u>Sand-silt-clay (cores)</u>					
	1.53	70.5	1602	Asian Shelf	3
<u>Silt (cores)</u>					
	1.60	60.6	1634	Japan Basin	3
	1.97	42.5	1650	NW Atlantic	1
	1.62	66.7	1560	Mediterranean	2
	1.55	69.8	1613	Mediterranean	2
	1.80	52.9	1598	NW Atlantic	1
	1.98	41.9	1691	NW Atlantic	1
<u>Clayey silt (cores)</u>					
	1.85	51.6	1597	Asian Shelf	3
	1.44	73.6	1554	Bering cont. rise	3
	1.56	66.0	1586	Bering cont. rise	3

TABLE 7. (Continued)

Sediment Type	Density (g/cc)	Porosity (%)	Velocity (m/sec)	Area	Ref.
<u>Clayey silt (cores) (continued)</u>					
	1.71	58.3	1543	Mediterranean	2
	1.74	57.7	1601	Mediterranean	2
<u>"Indurated" red clay (cores)</u>					
	1.23	87.8	1526	N. of Mid-Pac. Mtns.	4
	1.38	79.2	1548	Hawaiian Arch	4
	1.25	86.0	1536	Hawaiian Arch	4
	1.38	79.8	1543	Hawaiian Arch	4
<u>Mudstone (diver, submersible, dredge)</u>					
	1.84	52.3	1916	S. Calif. Shelf	3
	1.79	55.3	1576	La Jolla Canyon	3
	1.72	58.7	1593	La Jolla Fan	3
	1.58	67.1	1676	N. Wall Puerto Rico Tr.	3
<u>"Semi-consolidated" Cretaceous red clay and turbidites (Layer A)</u>					
			1680	N. Atlantic; cored in outcrop; velocities with sonobuoys	5.
			1830		6.
			2170		7
			2580		
			2850		
			2910		
<u>"Hard Mudstone" (Layer Beta)</u>					
			2900	N. Atlantic; see note above for Horizon A	5,
			2910		6,
			3470		7,
			3750		8
<u>Shale (oil-well log; top of section)</u>					
	2.10	(36.0)	1830	Gulf of Mexico	9

*References:

- | | |
|-------------------------------|----------------------------------|
| 1. Horn <i>et al.</i> , 1967b | 2. Horn <i>et al.</i> , 1967a |
| 3. Hamilton, this report | 4. Lair and Sanko, 1968 |
| 5. Ewing <i>et al.</i> , 1966 | 6. Saito <i>et al.</i> , 1966 |
| 7. Houtz <i>et al.</i> , 1968 | 8. Windisch <i>et al.</i> , 1968 |
| 9. Musgrave and Hicks, 1966 | |

TABLE 8. EXAMPLES OF CALCAREOUS SEDIMENTS AND ROCKS FROM THE SEA FLOOR.*

Sediment, Rock Type	Density (g/cc)	Porosity (%)	Velocity (m/sec)	Area	Ref.**
<u>Calcareous ooze (cores)</u>					
	1.38	79.0	1490	Atlantic	1
	1.47	72.8	1525	S. Pacific	2
	1.48	73.2	1493	NE Atlan. (avg. 2)	3
	1.54	69.2	1515	NE Atlan. (avg. 2)	3
	1.57	67.4	1556	NE Atlan. (avg. 2)	3
	1.59	66.5	1507	NW Atlan. (avg. 2)	4
	1.56	67.6	1504	NW Atlan. (avg. 9)	4
	1.59	66.4	1508	NW Atlan. (avg. 7)	4
	1.69	60.3	1524	NW Atlan. (avg. 7)	4
<u>Indurated (?) calcareous ooze (core)</u>					
	1.80	52.0	2370	Atlantic	5
<u>Limestone, reef; Eniwetok Atoll (drilling and seismic)</u>					
First layer			2440	Drilling and seismic	6
Second layer			3060		7
<u>Limestone, foraminiferal (dredge hauls)</u>					
	1.56	68.0	1770	Tertiary; guyots, Mid- Pac Mtns.; last 4 partially phosphatized	8
	1.73	62.0	1950		
	2.40	24.0	3860		
	2.64	12.0	4750		
	2.76	5.0	5460		

*Laboratory, 1 atmosphere pressure (except Eniwetok Atoll).

**References:

- | | |
|---|--------------------------------|
| 1. Schreiber, 1968a; representative value | 5. Sutton <i>et al.</i> , 1957 |
| 2. Hamilton, this report | 6. Ladd <i>et al.</i> , 1953 |
| 3. Horn <i>et al.</i> , 1967b | 7. Raitt, 1957 |
| 4. Horn <i>et al.</i> , 1968d | 8. Hamilton, 1959 |

TABLE 9. EXAMPLES OF SILICEOUS SEDIMENTS FROM THE SEA FLOOR, AND LABORATORY MEASUREMENTS ON ROCKS FORMED BY DIATOMS AND RADIOLARIA.*

Sediment, Rock Type	Density (g/cc)	Porosity (%)	Velocity (m/sec)	Area, Remarks	Ref.**
<u>Radiolarian ooze (cores)</u>					
	(1.20)	86.5	1520	Pacific (from fig. 3 in ref.)	1
	(1.21)	85.7	1488		
	(1.21)	85.5	1517		
	(1.29)	79.0	1470		
	(1.29)	79.0	1480		
	(1.29)	79.0	1500		
	(1.30)	78.0	1504		
	(1.34)	75.0	1492		
<u>Diatomaceous ooze (with radiolaria and spicules; cores)</u>					
	1.50	70.1	1644	Bering Sea	2
	1.41	75.3	1568	(avg. 3)	5
	1.37	76.9	1569	(avg. 5)	
<u>Diatomaceous "earth" (laboratory sample)</u>					
	(1.33)	75.0	1563	Saturated	3
<u>Radiolarian "earth" (laboratory sample)</u>					
	(1.35)	73.0	1657	Saturated	3
<u>Tripolite (diatomaceous earth; laboratory sample)</u>					
	(1.56)	57.0	2040	Sat; avg. 4	3
<u>Mixture basalt, chert, and siliceous sedimentary rock (seismic refraction)</u>					
			4200- 5200	N. Wall Puerto Rico Tr; correlated with dredge hauls	4

*Values in parentheses computed by writer.

**References:

1. Schreiber, 1964a
4. Rowin *et al.*, 1966

2. Boyce, 1967
5. Velocities (Hamilton, this report)

3. Wyllie *et al.*, 1958

TABLE 10. EXAMPLES OF LAYERS IN MARINE SEDIMENTS AS A RESULT OF VOLCANISM.*

Sediment Type	Density (g/cc)	Porosity (%)	Velocity (m/sec)	Area	Ref.**
Volcanic ash, glass, pumice					
Sand	1.98	48.0	1664	Nr. Hawaiian Ridge (avg. 4)	1
Silt	1.50	68.6	1634	Nr. Hawaiian Ridge (avg. 4)	1
Silt	1.30	84.7	1560	Nr. Hawaiian Ridge (avg. 4)	1
Silt	1.63	63.9	1562	Celebes Sea Basin	2
Silt	1.55	69.1	1580	Celebes Sea Basin	2
Silt	1.77	55.3	1620	Celebes Sea Basin	2
Sand	2.14	34.0	1832	Mediterranean	3
Silty sand	2.08	40.0	1713	Mediterranean	3
Sandy silt	1.76	56.7	1600	Mediterranean (avg. 3)	3
Silt	2.07	44.0	1674	Mediterranean	3
Silt	1.59	69.0	1587	Mediterranean (avg. 2)	3
Silty clay	1.88	54.0	1764	Mediterranean	3
Average	1.75	59.2	1634	20 samples	

*Laboratory, 1 atmosphere pressure.

**References:

1. Schreiber, 1968b
2. Hamilton, this report
3. Horn *et al.*, 1967a

Terrigenous Materials

Layers formed from terrigenous materials are those derived from sources on land (excluding volcanic materials) which are deposited in layers through the actions of turbidity currents, from sediment slumping, sliding and creeping, from eolian action, and ice rafting. These layers may be from a few millimeters to more than 1 meter in thickness.

The most common type of layer formed by terrigenous materials is deposited from turbidity currents moving along the sea floor. These deposits are called turbidites and are apt to be graded: that is, there is a gradation from coarse particles at the bottom to finer materials at the top. These layers are usually separated by finer silt-clays, deposited through this grading process, or from the deposition of fine particles through the water column (pelagic deposition). A thick turbidite section will thus be a continuous sequence of fine-grained, high-porosity, low-velocity layers, and thinner, coarser, lower-porosity, higher-velocity materials.

In general, turbidite layers, and those formed by other processes, can have the properties of the full range of sediments listed in table A-1 for continental-terrace sediments (sand to silty clay). In deep-water abyssal plains, the layers are usually formed by fine sand, sandy silt, silty sand, sand-silt-clay, and clayey silt. The finer materials are apt to be found at greater distances from source areas. Recent papers include discussions of turbidite layers in the North Pacific (Horn *et al.*, 1969a), and in the Tyrrhenian Abyssal Plain (Kermabon *et al.*, 1960).

In the upper levels of sediment represented by cores, layers are usually not lithified ("consolidated" in some geologic terminology). However, in the sea floor there is a full range from unlithified sediments to their lithified equivalents. Known examples are the sedimentary rocks composed of mineral particles from terrigenous sources, including mudstone, shale, limestone, and sandstone. Table 7 lists some typical examples, noted in the literature and from the writer's measurements.

Calcareous Materials

Calcareous ooze forms the second most common sediment type on the sea floor. In the open oceans, calcareous deposits are largely formed by the calcareous tests of Foraminifera and coccoliths. Near coral reefs the material is reef detritus. Limestone is formed when these calcareous particles become cemented (Hamilton, 1956, 1959; Friedman, 1964; Fischer and Garrison, 1967; Thompson *et al.*, 1968).

Sutton *et al.* (1957), Horn *et al.* (1968a), and Schreiber (1968a) have shown that calcareous oozes have little, if any, significant differences in sound velocity from deep-water silt clays of the same density and porosity; the few measurements by the writer are in accord. However, when limestone is formed, the velocity of compressional waves increases markedly. Examples are available from the sea floor of a full range from calcareous sediments to low-porosity, high-velocity limestones (table 8).

Siliceous Materials

The tests of Radiolaria and frustules of diatoms are formed of opaline silica. This material collects on the sea floor to form radiolarian and diatomaceous oozes. From marine deposits now found on continents, these deposits are known to form siliceous rocks (diatomite, tripolite, chert, etc.) through solution and redeposition of silica. In the sea floor, it has been demonstrated by the JOIDES drilling that siliceous rocks such as chert form hard layers within the sediment body (Peterson and Edgar, 1969). Such rocks have been dredged from the north wall of the Puerto Rico Trench (Bowin *et al.*, 1966; Bunce and Hersey, 1966). Properties of some siliceous sediments and rocks from the present-day or ancient, continental seas are listed in table 9.

Volcanic Materials

FRAGMENTAL MATERIALS

The most common layers in unlithified sediments of the deep sea are formed from fragmental, volcanic materials. The most common types are pyroclastic materials (those projected into the air by volcanoes). Macdonald (1967) recently discussed the various volcanic materials included in this classification. Horn *et al.* (1969a) and others have shown that volcanic ash and tuff are the most common type of pyroclastic materials which form layers in North Pacific sediments. The ash is usually composed of glass, cinders, mineral crystals and fragments, and other finer-sized materials. Most deep-sea ashes fall into the silt-size category, but a range from sand to silty-clay sizes is known. Horn *et al.* (1969a) have an excellent resumé of the extent and sources of ash layers in the North Pacific. Concerning thicknesses of these layers, Horn *et al.* analyzed 163 white ash layers (northwestern Pacific) which varied in thickness from 1 to 29 cm, with an average of 6.5 cm; brown ash (82 samples from the north-central Pacific) varied from 1 to 13 cm, with an average of 3.9 cm. Most of these layers were medium-grained silts.

Near volcanic islands, and ridges subject to erosion, there is a full range of sediment sizes (boulders to clayey silt) from volcanic sources. These volcanic sediment distributions in the Hawaiian area have been discussed by Hamilton (1957), Moberley and McCoy (1966), Schreiber (1968b), and Lair and Sanko (1968). Turbidites formed from ash, and sand-sized volcanic particles are common. The acoustic properties of representative examples of layers formed by volcanic, fragmental materials are listed in table 10.

BASALT LAYERS

Igneous rock (basalt) forms the upper oceanic crust and underlies the sediments and sedimentary rocks of the world's oceans. The basaltic rock of the upper crust has always been of considerable importance and interest in geophysics, but there are numerous, large areas of the sea floor where this rock, and lavas not connected with the crust, are close enough to (or at) the sea floor to be of importance in underwater acoustics.

The first hard, strongly reflective layer under the sediment layer in the deep-sea floor is often referred to as "Layer 2." Detailed seismic refraction and reflection surveys, coring of reflecting horizons, and other studies indicate that, in many areas, there are multiple layers in and above the oceanic crust, and the first hard, reflector ("Layer 2") under the soft sediments ("Layer 1") can be mudstone, shale, limestone, siliceous rock, or basalt. There can also be mixtures of these rocks and basalt, or basalt with soft sediments, as photographed on the East Pacific Rise (Bonatti, 1968). Thus, the term "Layer 2" loses its meaning with detailed work, and should probably be abandoned.

In many areas of the North Pacific, volcanic rock is either at the sea floor, or buried by a thin layer of sediments. For example, reflection profiling over the Hawaiian Deep and Arch have shown that the sediment cover is thin, from a few meters to non-existent, in many areas (Shor and Pollard, 1964; Normark and Shor, 1968; Kroenke, 1965). Unpublished reflection profiles taken by the NURDC, San Diego, in the Hawaiian Deep indicate 0 to 40 m of sediment over rock in several areas; in the northeast Pacific, between Erben and Fieberling Guyots, sediment thicknesses are no more than 30 to 40 m. In the Gulf of Alaska adjacent to hilly areas, and northwest of Guadalupe Island, there are large areas of flat, volcanic rock with little or no sediment cover (NURDC, unpublished). Low hills in the Gulf of Alaska have about 60 m of pelagic-sediment cover (Hamilton, 1967).

The basaltic layer underlying sediments in several areas in the Pacific has been traced by refraction and reflection surveys from the deep-sea floor into adjacent seamounts, ridges, rises, islands, or drilling (as at Eniwetok Atoll). Some examples are listed in table 11. Here, the velocity range (3.5 to 6.5 km/sec) shown for the upper crust is greater than might be expected considering the following factors: (1) the upper crust is considered to be a common type of rock (tholeiitic basalt); (2) basalts extruded in deep water should have porosities near zero and densities of the order of 2.9 to 3.1 g/cc; and (3) the velocity range in this material should be small (6.0 to 7.0 km/sec).

TABLE 11. EXAMPLES OF VELOCITIES IN "BASALT" LAYERS (NORTH PACIFIC AND ADJACENT AREAS).

Area	Velocity (km/sec)			Remarks	Ref.*
	Max.	Min.	Avg.		
Aleutian Islands	4.33	3.67	4.07	Al. Basin (5 stas.)	1
First basalt layer	-	-	3.19	Top of ridge	
Second basalt layer	-	-	3.81	Top of ridge	
Gulf of Alaska	-	-	5.45	Abyssal Plain	2
Juan de Fuca Ridge	6.04	4.99	5.35	Ridge (3 stas.)	3
Cascadia Basin	-	-	5.54	E. of ridge	3
Hawaiian Islands (Maui)	4.53	3.88	4.22	Arch-Deep (3 stas.)	4
	4.50	3.65	-	Ridge (2 stas.)	
Eniwetok Atoll	-	-	4.15	Seismic; drilling	5
Second basalt layer	-	-	5.59	Seismic	
Okhotsk Sea	-	-	6.50	Basin	6
Emperor Seamounts	3.93	3.56	3.75	Between smts.	7
East Pacific Rise	5.11	4.35	4.76	Top and W. side	8

*References:

- | | |
|------------------------------|--------------------------------|
| 1. Shor, 1964 | 5. Raitt, 1957 |
| 2. Shor, 1962 | 6. Koylin <i>et al.</i> , 1965 |
| 3. Shor <i>et al.</i> , 1968 | 7. Den <i>et al.</i> , 1969 |
| 4. Shor and Pollard, 1964 | 8. Raitt, 1956 |

Engel and Engel (1965) recently reviewed their own and other studies of crustal basalt and concluded that the upper oceanic crust is formed by tholeiitic basalt. The alkali-rich basalts which are common on the upper flanks and tops of seamounts, ridges, and islands are apparently derived from the primary, oceanic tholeiitic magma.

Basalt extruded in water depths of more than about 4000 m are apt to have little or no pore space in the form of gas-induced vesicles because of high hydrostatic pressures; the density of this material should be about 3.0 to 3.1 g/cc (Moore, 1965; Engel *et al.*, 1965). Figure 6 is reproduced from Moore's paper. These values are confirmed by the studies of Manghnani and Woollard (1968): tholeiitic basalts have average grain densities in the range of 3.0 to 3.2 (average: 3.05 g/cc; table 12), which would be the bulk densities of these rocks at, or near, zero porosities. If the postulates of sea-floor spreading are true, basalts extruded on top of a ridge or rise, and moved into deeper water, would be

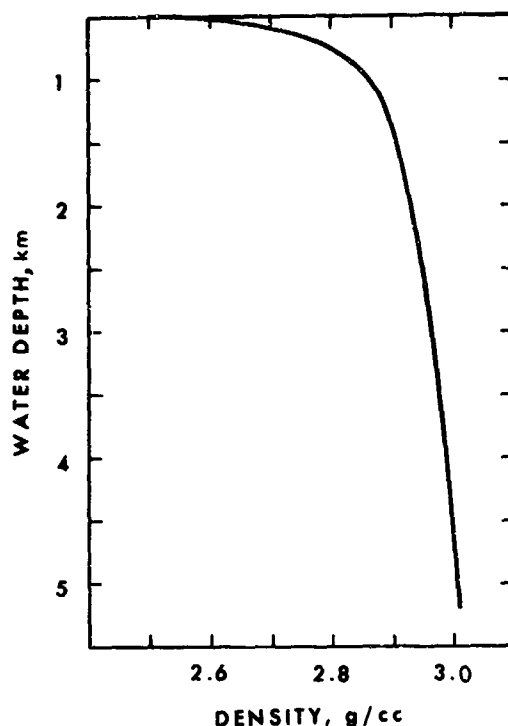


Figure 6. Density of basalt vs. water depth of magna extrusion (Moore, 1965).

more vesicular than if extruded at present water depths. The greater the vesicularity, the lower the density and velocity (Manghnani and Woollard, 1968). However, most centers of supposed spreading are deep enough that vesiculation would have been low; for example, basalts extruded on top of the East Pacific Rise, at about 3000 m, should have densities on the order of 2.96 g/cc, indicating a porosity of about 3 percent (Moore, 1965).

Basalts dredged from the deep-sea floor have high glass contents (Moore, 1965, Engel and Engel, 1965). Manghnani and Woollard (1965) have shown that velocity decreases with increasing glass content; and Manghnani *et al.* (1968) showed that velocity in obsidian decreases with increasing pressure. This velocity decrease although small ($V_p = 5.715 - 0.018P$; p in kilobars, velocity in km/sec) would contribute to lesser velocities in the oceanic crust.

Basalts show anisotropic velocity relationships depending on mineral and vesicle orientation, and porosity (Manghnani and Woollard, 1968; Balakrishna and Ramana, 1968). The anisotropic relationship is of the order of 2 to 3 percent for the higher-density basalts. A seismic-refraction, velocity measurement is parallel to the sea floor and normal to the overburden pressure, and would be apt to measure any such anisotropic relationship. Study of this anisotropic relationship, if it is present has not advanced enough to determine whether the greater velocity is parallel or normal to the sea floor. In sedimentary rocks the greater velocity is normal to overburden pressure and parallel with bedding planes.

Table 12 is a resumé from the excellent study of Hawaiian basalts recently completed by Manghnani and Woollard (1968). As these authors note, compressional-wave velocities in these basalts at 1 atmosphere pressure cannot be used to estimate velocities

TABLE 12. LABORATORY MEASUREMENTS (1 ATMOSPHERE PRESSURE) IN HAWAIIAN THOLEIITIC BASALTS AND OTHER SELECTED SAMPLES.

Rock Type (no. samples)	Grain Density (g/cc)	Density (g/cc)	Porosity (%)	V_p (m/sec)	V_s (m/sec)	$\frac{V_p}{V_s}$	Ref.*
Tholeiite (14)	3.06	2.60	14.5	4490	2590	1.73	1
Tholeiitic basalt (21)	3.00	2.58	14.1	4050	2530	1.60	1
Tholeiite olivine basalt (11)	3.05	2.44	20.0	3910	2370	1.65	1
Obsidian (1)	-	2.36	-	5715	3526	1.62	2
(1)	-	2.35	-	5820	3570	1.63	3
Deccan basalt (mean values)	-	2.93	-	6330	2900	2.18	4
Mohole basalt (1)	-	2.82	2.06	5760	2530	2.28	5
Basalt (1)	-	2.97	-	6480	3580	1.81	3

*References:

1. Manghnani and Woollard, 1968
2. Manghnani *et al.*, 1968; Pressure = 1 bar
3. Woeber *et al.*, 1963
4. Balakrishna and Ramana, 1968
5. Somerton *et al.*, 1963; dry density

in basalts in the sea floor under water and sediment hydrostatic and overburden pressures. The surface of the oceanic crustal layer at water depths greater than 4000 m is under water and sediment pressures on the order of 400 to 1000 bars. The temperature effect on velocity in the upper oceanic crust should be small (see resumé by Nafe and Drake, 1968).

Figure 7 illustrates density-velocity relationships (regression lines) for higher-density (>2.80 g/cc) Hawaiian basalts (data from Manghnani and Woollard, 1968). At zero porosity, the average density of tholeiitic basalts is about 3.05 g/cc; under a pressure of 1 kb, the velocity is about 6.83 km/sec. This is close to the velocity estimated by Woollard at 1 kb (1968: Curve B, 6.9 km/sec). At 0.5 kb the velocity at 3.05 g/cc is about 6.78 km/sec. Rounded to the nearest 0.1, the velocity at both 0.5 and 1 kb is 6.8 km/sec.

In areas where the first strong reflector is thought to be basalt, the low velocities frequently measured (3.5 to 5.5 km/sec) are unlikely to represent basalt of varying porosities because most of these basalts were extruded in deep water where their porosities should be near zero. In this case, the low velocities cannot be related to the density-velocity relationships which have been established for basalts and other rocks. For example, off Hawaii an upper crustal velocity of 4.2 km/sec (Shor and Pollard, 1964) would indicate a density of about 2.3 g/cc (Manghnani and Woollard, 1968). Given a value of 3.05 g/cc as the bulk density of minerals in tholeiitic basalt, a density of 2.3 g/cc requires a porosity (vesicularity) of about 25 percent, which is unlikely because of the probable water-depth of extrusion of the basaltic magma.

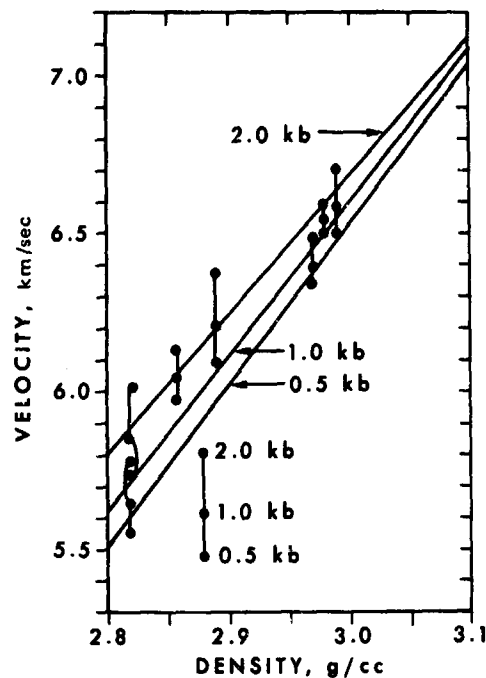


Figure 7. Density vs. velocity in higher-density (>2.8 g/cc) Hawaiian basalts at various pressures. Data from Manghnani and Woollard (1968).

As noted by Strange *et al.* (1965), Nafe and Drake (1967), Nayudu (1969), and others, it is more likely that the upper surface of the oceanic crust is composed of basalt in the form of flows, pillows, and fragmental material, with sedimentary material filling cracks, fissures, and areas between pillows and flows. Photographs of the sea floor on the East Pacific Rise (Bonatti, 1968), dredge hauls off Hawaii (Moore, 1965), and studies of magma intruded into "deep-water" sediments such as those now exposed on Unalaska Island (Snyder and Fraser, 1963; Moore and Fiske, 1969) indicate that subaqueous basaltic magmas are apt to form pillows with sediment intercalated between them. In the formations on Unalaska, the matrix between pillows is pepperite, a mixture of fragmental volcanic material and sediment.

A refraction measurement of compressional-wave velocity is along the top of a layer and parallel to the sea floor, and is not the interval velocity in the layer. Thus a refraction measurement along the top of the oceanic crust should "integrate" or average the velocities of all materials along the ray path in the same way as indicated by a "normalized velocity or interval velocity curve" for a well log in a section composed of alternating layers of shale and sandstone (*e.g.*, Musgrave and Hicks, 1966, figs. 5, 7, 8). If the upper oceanic crust is composed, in many areas, of flows and pillows of solid basalt with sediment and fragmental material filling cracks, fissures, and "pore spaces," the velocity through this section should be directly related to velocity in the solid basalt and in the intercalated material.

Given a refraction measurement of velocity along the top of a layer composed of basalt and sedimentary material (sediment and sedimentary rock), it should be possible

to compute the relative amounts of each material in the section, if reasonable values for velocity in each material can be assumed. This method employs the equation

$$V_1 t_1 + V_2 t_2 = V_3 t_3$$

where

V_1 and t_1 = velocity and time in basalt

V_2 and t_2 = velocity and time in sedimentary material

V_3 and t_3 = measured velocity and time (seismic refraction) in the section

$t_1 + t_2 = t_3$; let $t_3 = 1$ sec; $t_1 + t_2 = 1$, and $t_2 = (1 - t_1)$

then

$$V_1 t_1 + V_2 (1 - t_1) = V_3; t_1 = \frac{V_3 - V_2}{V_1 - V_2}$$

Distance traveled in (or relative amount of) basalt = $V_1 t_1$

Distance traveled in (or relative amount of) sediment = $V_2 t_2 = V_2 (1 - t_1)$

Table 13 has been computed by this method, plus the assumptions:

1. The layer is a mixture of basalt, basalt fragments, sediments, and sedimentary rock at a water depth of 5000 m; no sediment cover;
2. Properties are: basalt velocity and density are 6.8 km/sec and 3.05 g/cc, sediment velocity and density are 1.7 km/sec and 1.8 g/cc, and section velocities from seismic refraction measurements are known (first column).

TABLE 13. RELATIVE AMOUNTS AND PROPERTIES OF BASALT AND SEDIMENTARY MATERIALS IN BASALT-SEDIMENT SECTIONS.*

Compressional Velocity, V_p (km/sec)	Amount (%)		Density, Section (g/cc)	Shear Velocity, V_s (km/sec)
	Basalt	Sediment		
3.50	69	31	2.66	2.02
3.80	74	26	2.72	2.20
4.00	77	23	2.76	2.31
4.20	79	21	2.79	2.43
4.50	83	17	2.84	2.60
5.00	88	12	2.90	2.89
5.50	92	8	2.95	3.18

*See text for method of computation and assumptions.

Given the relative amounts of basalt and sedimentary material, the density of the section can be computed (table 13) with assumed values (as above) for density in each material: average density = volume of basalt (density of basalt) + volume of sediment (density of sediment).

The shear-wave velocities in table 13 were computed with a representative ratio: $V_p/V_s = 1.73$, which has been reported for the top of the crust in the Atlantic (Houtz and Ewing, 1963), and in Cascadia Basin (Shor *et al.*, 1968).

It should be reiterated that velocities in the range considered (3.5 to 5.5 km/sec) can also represent sedimentary rock, and that the preceding reasoning and computations apply only to areas where the layer in question is known to be mostly "basaltic"; for example, on the East Pacific Rise, the Mid-Atlantic Ridge, or where the rock layer under a layer of sediment has been traced to adjacent hills or ridges.

Crustal Densities

Assuming that the previously developed line of reasoning and computations are reasonable generalizations, the density of crustal sections and the whole crust can be computed, given velocity measurements and thicknesses of the layers, and assuming reasonable velocities and densities for the basalt and sedimentary material. The implied assumption is that the oceanic crust is composed of basalt flows and pillows and intercalated sedimentary materials at the top, increasing amounts of basalt relative to sedimentary material at deeper levels, and of solid basalt in the lower levels (with all layers under increasing pressures). It is interesting to note that the velocity value selected for solid basalt between 0.5 and 1 kb (6.8 km/sec) is a commonly measured velocity value for "Layer 3," the bottom layer of the oceanic crust; small, measured variations in these velocities, of the order of a few tenths of a km/sec, may represent small compositional variations and/or responses to varying overlying pressures. This reasoning and the computations in no way deny the possibility that the lower oceanic crust is serpentinized peridotite (Hess, 1955, 1964) because the velocities in this type of material at 1 and 2 kb (Birch, 1961, 1964; Simmons, 1964) are too close to the velocities in basalt to differentiate the rock type.

Some oceanic crustal densities were computed by using the method discussed in the previous section for each layer, and then summing the layer contribution to crustal density from its percent of total, crustal thickness. Assumed basalt properties were 6.8 km/sec and 3.05 g/cc, and assumed, intercalated sedimentary properties were 2.2 km/sec and 2.0 g/cc. Results were:

1. Hawaiian Arch (Sta. 29, Shor and Pollard, 1964)2.96 g/cc
2. East Pacific Rise (Sta. C-20, Raitt, 1956)2.99
3. Mid-Atlantic Ridge (Talwani *et al.*, 1965); including sediments at appropriate velocities and densities:
 - a. West flank
 - Sta. A180-12.88
 - Sta. A180-22.78
 - b. East flank
 - Sta. V10-82.91
 - Sta. V10-92.90

All these oceanic crustal densities are greater than currently accepted and used in these areas for gravity studies.

As previously noted, a seismic refraction measurement of velocity is at the top of a layer and is not the interval velocity. There is undoubtedly a positive velocity gradient in the upper oceanic crust because of pressure. To use the surface velocity as the interval velocity results in computations of layer thickness which are too small. However, when a depth in the crust is reached where there is igneous rock at zero porosity, the increase of velocity with expectable pressures is probably small (Simmons and Nur, 1969).

The author offers this type of computation of crustal density merely as an interesting speculation; there are too many imponderables to defend it. However, there seems a reasonable possibility that the uppermost surface of the crust is composed of pillows and flows of basalt at, on, near, zero porosity, with intercalated sediments and sedimentary rocks—in which case, computations of relative amounts of basalt and sedimentary materials, and average densities, may be reasonable.

VELOCITY GRADIENTS

Introduction

The emphasis in this report has been on velocities and related properties in the sediment surface. The geophysicist and underwater acoustician require the gradients of velocity with depth in the sediment. All sediment and rock layers have such gradients, and it would be incorrect to assume that a layer of any appreciable thickness is isovelocity with depth; it would also be incorrect to use the sediment surface velocity to compute ray paths, true thicknesses of layers, and other factors involving sound propagation within layers.

It is possible to approximate velocity changes with depth in the sea floor by corrections to surface velocity caused by overburden pressure, reduction of porosity, temperature increases with depth, and a number of other complex and interrelated factors (Hamilton, 1959, 1964). However, at best, these approximations are always tenuous. It is much better to use layer interval velocities and velocity gradients measured *in situ* in the sea floor. At present, the simplest method for such measurements involves the use of expendable sonobuoys.

The relatively new technique of velocity measurements with expendable sonobuoys has been described by Le Pichon *et al.* (1968), and will not be reviewed here. The results of sonobuoy measurements in the Atlantic and adjacent areas, and in the central Pacific, have been recently summarized in excellent papers by Houtz *et al.* (1968) and Ewing *et al.* (1969). Some of these results are listed in table 14, illustrated in figure 8, and discussed below. The expendable-sonobuoy technique has been used by Scripps Institution and NURDC (reports in preparation).

Velocity gradients are usually expressed as an increase in velocity per linear increase in depth, or sec^{-1} . In upper levels of deep-water marine sediments these gradients are positive, and usually between 0.5 and 2.0 sec^{-1} (Ewing and Nafe, 1963).

In the discussion in this section, several velocities are involved: (1) instantaneous velocity, V , is the velocity of a compressional wave at any given depth or travel time within the sediment body (hereinafter, also referred to as "velocity"), (2) mean velocity, or

TABLE 14. LINEAR VELOCITY GRADIENTS TO INDICATED SEDIMENT DEPTHS.

Area	Gradient (sec ⁻¹)	Sediment Depth (m)	Remarks
<u>Atlantic (Houtz <i>et al.</i>, 1968); sonobuoys</u>			
	1.291	0	From parabolic regression equations for sonobuoy data: this report, eqns. 11, 12, 15, and 16.
	1.153	78	
	1.031	160	
	0.850	334	
	0.733	517	
<u>Gulf of Mexico (Houtz <i>et al.</i>, 1968); sonobuoys</u>			
	1.134	0	From parabolic regression equations for sonobuoy data: This report, eqns. 13, 14, 15, and 16.
	1.089	79	
	1.031	163	
	0.948	345	
	0.877	543	
<u>Equatorial Pacific (Ewing <i>et al.</i>, 1969); sonobuoys</u>			
	(1.85)	(168)	Estimated by author.
<u>Gulf of Mexico (Musgrave and Hicks, 1966); well logs</u>			
Shale section	(0.6)	(152)	Computed by author from figures 5 and 10 in ref.
Shale and sandstone	(1.0)	(152)	

interval velocity, \bar{V} , is the average velocity for an interval or layer, and (3) sediment-surface velocity, V_0 , is compressional-wave velocity in the sediment just below the water-sediment interface. V_0 is plotted on diagrams of velocity vs. depth or time, at depth or time equal to zero.

There are several ways to construct diagrams showing the variations of these velocities with depth or time: (1) instantaneous velocity vs. reflection time (two-way travel time within the sediment), (2) instantaneous velocity vs. one-way travel time, (3) instantaneous velocity vs. depth within the sediment body, (4) interval (mean) velocity vs. one- or two-way travel time, or (5) interval velocity vs. depth. The curves in figure 8 are from Houtz *et al.*, (1968) and Ewing *et al.* (1969) and are plotted with one-way travel time for the reasons discussed in those reports.

In reducing sonobuoy data, the interval velocity, \bar{V} , is derived for one or more layers; after the true thicknesses, h , of the layers have been computed ($h = \bar{V}t$; t = one-way travel time), a plot of instantaneous velocity (velocity at any given depth or time) can be constructed by plotting the interval velocity at a depth (or time) equal to one-half the depth interval or time. Methods of construction and corrections to these plots were discussed by Houtz *et al.* (1968).

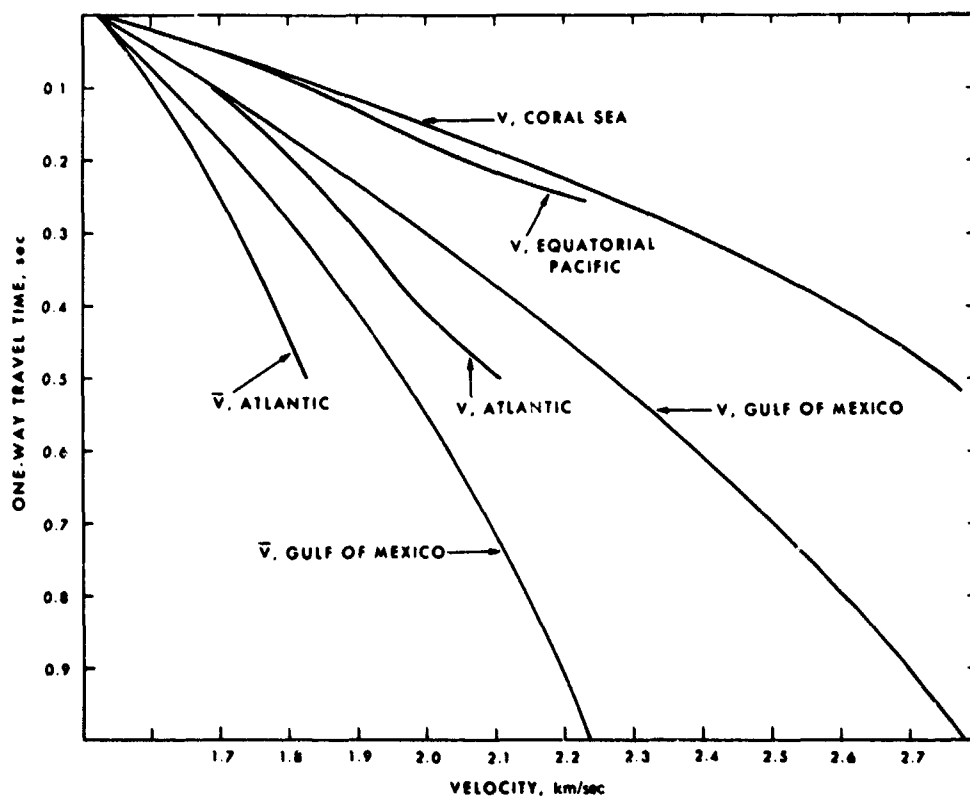


Figure 8. Instantaneous velocity, V , and mean (interval) velocity, \bar{V} , vs. one-way travel time, t , in Atlantic and Gulf of Mexico sediments; V vs. t on Coral Sea and Equatorial Pacific sediments (Houtz *et al.*, 1968; Ewing *et al.*, 1969).

One-way travel time in a sediment or rock layer can be measured directly from a reflection record. The recorder is usually set to read in fathoms at a water speed of 1463 m/sec (4800 ft/sec). Two methods of measuring one-way travel time are: (1) measure, from the record, the apparent thickness of the layer in fathoms and convert fathoms to time (seconds) by the relationship: 1 fathom = 0.00125 second, or (2) note on the top and bottom of the record the one-way travel time in seconds, and measure time directly.

In constructing generalized curves of velocity vs. depth or time, using sonobuoy data, the problem arises as to what velocity should be used at time or depth equal to zero, or, expressed as a question: what is a good, average value of the velocity at the sediment surface? This problem occurs because the first interval velocity derived from sonobuoy data is at the midpoint of the first layer. Houtz and Ewing (1964), Houtz *et al.* (1968), and Ewing *et al.* (1969) used 1520 m/sec for the sediment-surface velocity. This value was selected as the result of wide-angle reflection measurements in the Atlantic at three stations where the water depth averaged 5380 m (Houtz and Ewing, 1964). This value (1520 m/sec) is also a good general choice for much of the Pacific where about 62 percent

of the sea floor is at water depths between 4000 and 6000 m (Menard and Smith, 1966). The bottom-water velocity in the Central Pacific at 5000 m is about 1542 m/sec. A sediment-surface velocity of 1519 m/sec is derived by using the ratio, velocity in sediment/velocity in bottom water, for abyssal-hill pelagic sediment (0.985; table A-2). However, the sediment-surface velocity is closely linked to the bottom-water velocity through the velocity ratio (discussed in the section "Velocity"), and the bottom-water velocity varies with water depth. The amount of these variations can be approximated for abyssal silt-clays in the Central Pacific by multiplying the average velocity ratio (0.985) by the bottom-water velocity (table 1): at 3000-m water depth, a sediment-surface velocity of 1485 m/sec (rounded to the nearest 5 m/sec) should be used; at 4000 m, 1500 m/sec; at 5000 m, 1520 m/sec; at 6000 m, 1535 m/sec. Thus, for given regions the generalized curve of velocity vs. depth or time should pass through different sediment-surface velocities (at depth or time equals zero), depending on the average ratio and average depth of water in the area.

In areas of turbidite deposition, the smooth, velocity curves (e.g., fig. 8) are "integrating" or averaging many abrupt changes of velocity in low-velocity, silt-clay layers and higher-velocity, sand-silt layers. These smoothed curves are of chief utility in deriving the true thicknesses of sediment or rock layers in which true reflection time has been measured by reflection surveys. For several purposes of underwater acoustics, more detailed curves (e.g., from velocity measurements in a core), which show departures from these generalized curves, are necessary in turbidite sections.

Nonlinear Gradients

Most velocity gradients are nonlinear if followed to sufficient depths within the sediment body (fig. 8). Houtz *et al.* (1968) described the gradients in the Atlantic and the Gulf of Mexico with polynomial equations of the type: velocity, $V = A + Bt + Ct^2 + Dt^3$, where t is one-way travel time within the sediments. These curves are regression equations which fit the corrected interval-velocities of a large number of sonobuoy measurements.

The curves of velocity vs. one-way travel time (fig. 8) are composite curves for four general areas. For the Atlantic and Gulf of Mexico (Houtz *et al.*, 1968), both instantaneous velocity, and interval or mean velocity vs. one-way travel time are shown for, mostly, terrigenous sediments. The curves for the Coral Sea and Equatorial Pacific (Ewing *et al.*, 1969) represent instantaneous velocities in calcareous sediments. No equations were given for the Pacific curves (a more detailed paper is in preparation). The equations for the Atlantic curves (Houtz *et al.*, 1968, p. 2636, table 2) are

$$V = 1.5199 + 1.9628t - 3.4692t^2 + 3.7864t^3 \quad (11)$$

$$\bar{V} = 1.5199 + 0.8895t - 0.8781t^2 + 0.6318t^3 \quad (12)$$

The equations for the Gulf of Mexico curves (Houtz *et al.*, 1968) are

$$V = 1.5207 + 1.7241t - 0.4679t^2 - 0.0002t^3 \quad (13)$$

$$\bar{V} = 1.5202 + 1.1072t - 0.4906t^2 + 0.0983t^3 \quad (14)$$

where

V = instantaneous velocity, km/sec

\bar{V} = interval or mean velocity, km/sec

t = one-way travel time, sec

The thickness of any sediment layer, h , in kilometers, is

$$h = \bar{V}t \quad (15)$$

With these equations, plots of instantaneous or mean velocity vs. depth, and of thickness vs. time, can be constructed.

Linear Gradients

In the upper parts of sediment layers, the water-sediment interface velocity gradient for nonlinear curves ($t = 0$) can be used as a linear gradient to shallow depths with little error (examples below). Serious errors can occur in thickness computations and instantaneous velocities at greater depths unless parabolic curves or equations are used.

At any depth (or any one-way reflection time) within sediment layers, an average linear gradient can be determined from the parabolic equations for an established curve (Houtz *et al.*, 1968, eq. 3) by

$$\text{average gradient} = (V - V_0)/h \quad (16)$$

where

V = instantaneous velocity at time t

V_0 = velocity at sediment surface ($t = 0$)

h = layer thickness at time t

Table 14 lists gradients at $t = 0$, and average linear gradients to stated depths, from the nonlinear curves for the Atlantic and Gulf of Mexico (Houtz *et al.*, 1968).

If only one sonobuoy determination of interval velocity in one layer is available, a linear gradient, a , can be found by

$$a = \frac{(\bar{V} - V_0)}{h/2} \quad (17)$$

where

\bar{V} = mean or interval velocity

V_0 = *in situ* sediment surface velocity determined as discussed in section "Velocity"

h = layer thickness determined by: $h = \bar{V}t$; where t equals one-way travel time in the section.

As an example of the construction of a linear gradient from a one-layer interval velocity determination, a gradient was computed for Pacific pelagic sediments (probably calcareous) in an area southeast of Hawaii (Houtz *et al.*, 1968, table 1, fig. 12, V24-25).

The information may be tabulated as follows.

Station	Upper Reflection Time (sec)	Lower Reflection Time (sec)	Interval Velocity (m/sec)	Water Layer Reflection Time (sec)
V24-25	6.059	6.577	1807	6.059

From which were derived:

1. one-way travel time in the layer: $t = (6.577 - 6.059)/2 = 0.259$ sec
2. thickness of layer, $h = \bar{V}t = 468$ m
3. true water depth:
 - a. one-way travel time = $6.059/2 = 3.0295$ sec
 - b. echo-sounder depth = 3.0295 sec (1463 m/sec) = 4432 m
 - c. water interval velocity: from Mathews Tables (NAVOCEANO 1966) = 1502 m/sec
 - d. true water depth = t (water interval velocity) = 4550 m
4. *in situ* sediment surface velocity, V_0 (see discussion in section "Velocity"):
 - a. bottom-water velocity at true water depth = 1534 m/sec
 - b. sediment surface velocity, $V = 0.985$ (1534) = 1511 m/sec

The linear velocity gradient at this section was then determined with equation 17:

$$a = (1807 - 1511)/234 = 1.265 \text{ sec}^{-1}$$

If the interval velocity for more than one layer has been determined, a single curve of velocity vs. depth or time can be constructed using the procedure outlined by Houtz *et al.* (1968). It may or may not be linear.

Given a linear gradient, a , the sediment surface velocity, V_0 , and one-way travel time, t , the thickness of a layer can be computed (Houtz and Ewing, 1963) by

$$h = V_0(e^{at} - 1)/a \quad (18)$$

This is a very useful equation because V_0 can be closely estimated, and one-way travel time in a layer can be measured from a reflection record; and, as discussed in this section, the velocity gradient can usually be reasonably estimated.

When the velocity gradient is linear, the instantaneous velocity, V , at depth, h , is (Houtz and Ewing, 1963):

$$V = V_0 + ah \quad (19)$$

Some linear gradients for various areas are summarized in table 14.

There is a tendency in marine geophysical literature to assume a value for the interval velocity in deep-sea sediments which is usually too high. Common assumptions are 2000 m/sec (which is convenient in computations) or, for the Pacific, 2150 m/sec, which was an old, average value based on "unpublished estimates of velocities determined from seismic reflection profiles" in the Pacific (Raitt, 1956). Now that interval velocity and sediment-surface velocity information is available, this type of assumption should be abandoned in favor of actual measurements or more realistic estimates and assumptions. For example (in the case of a measurement), at Station V24-25 (Houtz *et al.* 1968), discussed above, one-way travel time, t , in the pelagic sediment layer was 0.259 sec. Assuming that sediment interval velocity is 2150 m/sec yields a sediment thickness ($h = \bar{V}t$) of 557 m; if 2000 m/sec is assumed, $h = 518$ m. The correct value of interval velocity is 1807 m/sec, and the thickness is 468 m. The least difference between assumptions and measurement is 50 m.

At the present time there is a large amount of acoustic reflection data from which one-way travel time in sediment layers can be measured. Given one-way travel time, interval velocities (or velocity gradients and sediment surface velocities) are required to compute true thicknesses of layers; and interval velocities are usually not available. Rather than assuming interval velocities (such as 2000 or 2150 m/sec), it is better to

1. estimate a sediment surface velocity using the methods of this report (section, "Velocity"),
2. assume a velocity gradient based on that for similar sediments elsewhere, and
3. compute thickness, h , with equation 18.

As an example of the "error" in thickness involved in the alternative methods, above, assume:

1. that a layer is abyssal-hill pelagic clay in the Central Pacific,
2. a measurement of one-way travel time, $t = 0.1$ sec, in the sediment layer, and
3. true water depth = 5000 m.

If 2150 m/sec is assumed as interval velocity, \bar{V} , the thickness of the layer ($h = \bar{V}t$) is 215 m.

Using the recommended procedure:

1. Determine the velocity in the bottom water; central Pacific at 5000 m: 1542 m/sec,
2. Sediment surface velocity, $V_0 = 0.985(1542) = 1519$ (round to 1520 m/sec),
3. Assume the velocity gradient is 1.0 sec^{-1} (discussion below), and
4. Layer thickness, $h = V_0(e^{at} - 1)/a$ (eq. 18) = 160 m; if a gradient of 1.3 sec^{-1} (surface gradient Atlantic, table 14) was assumed, $h = 162$ m.

The difference between the thicknesses computed by assuming an interval velocity of 2150 m/sec (215 m) and that resulting from more realistic estimates (about 160 m), is about 55 m; which is a significant difference in studies of rates of sedimentation, and other aspects of geologic history.

Serious errors in computed layer thicknesses result if a surface, linear-velocity gradient is used for thick sections (high values of t), because the velocity vs. time or depth curves are parabolic and gradients decrease with increasing depth in the sediments (table 14; fig. 8). For example, in the Gulf of Mexico, in a section in which one-way travel time

is 1.0 sec, the "true" thickness using the equation for the parabolic curve for mean velocity, \bar{V} , is 2235 m. A thickness of 2826 m is computed if the surface gradient at $t = 0$ (1.134 sec^{-1}) is used in the linear equation (18).

Selection of Velocity Gradients

For an area in which no velocity-gradient information is available, the problem is to estimate a reasonable gradient based on gradients in similar sediments elsewhere. The unpublished interval velocities and velocity gradients determined from NURDC and Scripps Institution sonobuoy measurements in the Pacific (reports in preparation) are in good agreement with the velocity data published by the Lamont-Doherty group (Houtz *et al.*, 1968; Ewing *et al.*, 1969). Consequently, until the results of numerous additional measurements at the three institutions are available, the following velocity gradients (or estimated gradients) are recommended for use in the Pacific and adjacent areas.

TERRIGENOUS SEDIMENTS

In terrigenous sediments (including turbidites) it is recommended that the curves and equations established for the Atlantic and Gulf of Mexico be used for the Pacific. For areas near the base of the continental slope use the data from the Gulf of Mexico. For deep-sea areas farther away, use the Atlantic data (to about $t = 0.5$ sec). For thick sections ($t > 0.5$ sec) use the parabolic equations for the Gulf of Mexico. In some general studies, or where layers are relatively thin, and in some acoustic studies involving the upper tens of meters of sediments, it may be desirable or convenient to use a linear gradient. Inspection of the equations and curves for the Atlantic and Gulf of Mexico indicates that a gradient of 1.0 sec^{-1} can be used with negligible error to depths of about 160 m, or one-way travel time of 0.1 sec. Other, average linear gradients in Atlantic and Gulf of Mexico sediments, at given one-way travel times are listed in table 14.

CALCAREOUS SEDIMENTS

The curves of figure 8 from the Coral Sea and Equatorial Pacific (Ewing *et al.*, 1969) were based on measurements in dominantly calcareous sediments. For other areas presumed to be calcareous, the Equatorial Pacific curve can be used to about $t = 0.25$ sec. At greater values of t , the Coral Sea curve can be used (to about $t = 0.5$ sec). Linear velocity gradients can be approximated (pending publication of mean-velocity and instantaneous-velocity regression equations) by

$$a = (V - V_0)/\bar{V}t$$

where

V = instantaneous velocity at time t

V_0 = surface velocity (given: 1520 m/sec)

$$\bar{V} = (V + V_0)/2$$

t = one-way travel time in the layer

In general, a gradient of 1.9 sec^{-1} can be used to about $t = 0.1 \text{ sec}$ for calcareous sediments, and 1.5 sec^{-1} to about $t = 0.2 \text{ sec}$. At $t > 0.2 \text{ sec}$, where there is no indication of high-velocity layers at depth, the Coral Sea curve should be used.

PELAGIC CLAY

Data are insufficient to establish velocity gradients in deep-sea pelagic clay in the Pacific. Many of the measurements in the Atlantic were in this type of sediment; therefore, pending further studies, a linear gradient of 1.0 sec^{-1} can be used for these relatively thin layers. A sediment thickness chart of the North Pacific indicates that the sediment layer in the central basin (excluding the relatively thick calcareous sediments of the equatorial area) has a two-way reflection time of less than 0.1 sec (Ewing *et al.*, 1969).

Sound Channels in Surficial Deep-Sea Sediments

During studies of seismic-refraction measurements in the Atlantic, Hersey *et al.* (1952), Officer (1955), and Katz and Ewing (1956) noted a constant-frequency arrival at long ranges which was interpreted as refracted along the ocean floor, possibly within a layer of low-velocity sediment. Katz and Ewing received these constant-frequency arrivals between 30 and 100 Hz at ranges of 40 to 50 seconds (59 to 74 km) and at a travel time of the first reflection. Subsequent information on *in situ* sediment surface velocities and gradients independently verify that such sound channels should be present over very large areas in most of the world's oceans.

As noted in preceding sections, compressional-wave velocities in surficial deep-water silt-clays are typically less than in the water just above the sea floor (e.g., 1-1/2 percent less in abyssal-hill silty clay). This phenomenon, together with the velocity gradients noted in the preceding section, produces a small sound channel at the top of the sediment section.

Assuming average values of bottom-water and sediment velocities, and an expectable, average range of velocity gradients, figure 9 indicates the expectable dimensions of typical sound channels at water depths of 5000 m in the central Pacific in an area of homogeneous silty clay. The height of the sound channel varies from about 15 to 46 m, as the sediment velocity gradient varies between 1.5 and 0.5 sec^{-1} . These thicknesses compare with those estimated for the Western Atlantic by Katz and Ewing (1956): 5 to 50 meters, and those estimated at the Guadalupe Preliminary Mohole Site: 20 to 45 meters (Hamilton, 1965).

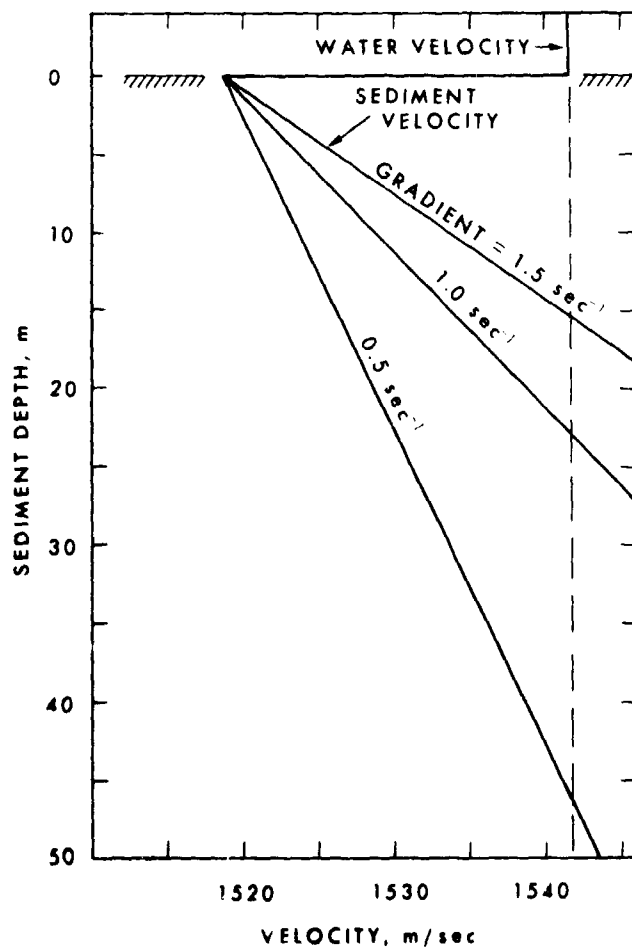


Figure 9. Diagram of sound channels expectable in Central Pacific surficial, deep-sea clay at various sediment velocity gradients.

APPENDIX: SUPPORTING DATA FOR USE IN PREDICTION OF *IN SITU* PROPERTIES

Parts I and II of this report included some tabulated data which are essential in the prediction of *in situ* properties. As it is intended to make each of the three parts self-contained, some of these data are reproduced here. Tables A-1 and A-2 appeared in TP 143; tables A-3 and A-4 are from TP144.

Table A-5 lists hydrostatic seawater pressures at various depths, temperatures, and salinities (from Wilson, 1959). Such tables are not commonly found in the literature. Because of local interest, table A-6 lists various water properties with depth in the San Diego Trough (about 15 nmi west of San Diego). Although surface-water properties vary seasonally, the properties listed at deeper depths (below about 150 m) remain relatively stable.

TABLE A-1: SEDIMENT PROPERTIES, CONTINENTAL TERRACE (SHELF AND SLOPE) ENVIRONMENT

Sediment Type	N Sample	Grain Diameter (mm)			C ₆₀ (%)	C ₂₅ (%)	Bulk Grain Density (g/cc)		Porosity (%)		Velocity (m/sec)		Ratio		p ₂ /p ₁		p ₂ /p ₃		R		BI	
		Mean (mm)	Median (mm)	Standard Deviation (mm)			Avg	SE	Avg	SE	Avg	SE	Avg	SE	Avg	SE	Avg	SE	Avg	SE	Avg	SE
Sand																						
	Coarse	1	0.580	0.520	100.0			2.71	2.03	38.6	...	1836	...	1.201	...	3.7347	...	6.8577	...	0.4098	...	7.8
Medium																						
	12	0.156	0.156	99.8	0.2		2.70	2.01	39.7	0.46	1749	6	1.144	0.004	3.5087	0.020	6.1303	0.050	0.3835	0.002	6.3	0.05
Fine																						
	9	0.153	0.151	88.1	6.3	3.1	2.70	1.98	0.024	43.9	1.29	1742	10	1.139	0.006	3.4433	0.040	6.0868	0.117	0.3749	0.005	5.6
Silt/clay																						
	3	0.090	0.090	83.9	11.0	2.9	2.74	1.91	...	47.4	...	1711	...	1.121	...	3.2645	...	5.5946	...	0.3517	...	9.1
Silt/clay																						
	11	0.074	0.126	65.0	21.6	13.4	2.71	1.83	0.025	52.5	1.55	1677	9	1.096	0.006	3.9633	0.050	5.1387	0.102	0.3228	0.008	9.9
Silt/clay																						
	6	0.056	0.051	14.5	41.2	14.3	2.75	1.56	...	68.3	...	1552	...	1.015	...	2.4201	...	3.7587	...	0.2136	...	13.5
Silt/clay																						
	17	0.019	0.041	32.6	41.2	26.1	2.71	1.58	0.030	67.5	1.66	1578	9	1.032	0.006	2.4939	0.059	3.9420	0.113	0.2504	0.010	12.1
Clay/silt																						
	20	0.006	0.031	6.1	69.2	14.8	2.71*	1.43	0.016	75.0	0.87	1535	3	1.004	0.002	2.1989	0.026	3.3782	0.045	0.1767	0.012	15.2
Clay/silt																						
	7	0.003	0.004	5.3	41.5	53.6	2.69	1.42	0.013	76.0	0.74	1519	3	0.994	0.002	2.1571	0.024	3.2804	0.042	0.1586	0.005	16.1

Notes: Laboratory value: 21.1 atmosphere
 Density: laboratory bulk density, presents all free ratio values on sediment velocity in water at 21°C, 1 atmosphere, and salinity of sediment pore water
 SE: standard error of the mean

*N/A: sediment impedance (cm/sec)² 10³

p_1/p_2 = sediment velocity ratio (cm/sec)² = $\frac{p_1 V_1}{p_2 V_2}$

R = Backscatter coefficient at normal incidence = $\frac{p_2 V_2 - p_1 V_1}{p_2 V_2 + p_1 V_1}$

RI = p_1/p_2 bottom on dB

p_1, p_2 = water density velocity p_1, p_2 = sediment density velocity

TABLE A-3: CONTINENTAL TERRACE (SHELF AND SLOPE) ENVIRONMENT: COMPUTED ELASTIC CONSTANTS IN SEDIMENTS

Sediment Type	κ		μ		λ		σ		V_s	
	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE
Sand										
Coarse	6.6859	---	0.1289	---	6.6000	---	0.491	---	250	---
Fine	5.6877	0.195	0.3212	0.064	5.3044	0.247	0.469	0.007	382	45
Very Fine	5.1182	---	0.5035	---	4.7825	---	0.453	---	503	---
Silty sand	4.6812	0.145	0.3926	0.051	4.3330	0.152	0.457	0.006	457	31
Sandy silt	3.4152	---	0.2809	---	3.2279	---	0.461	---	379	---
Sand-silt-clay	3.5781	0.095	0.2731	0.023	3.3995	0.090	0.463	0.003	409	18
Clayey silt	3.1720	0.040	0.1427	0.011	3.0735	0.040	0.478	0.002	364	57
Silty clay	3.1476	0.037	0.1323	0.029	3.0592	0.027	0.480	0.004	287	22

Notes: Laboratory values: 23°C, 1 atmosphere pressure

κ = bulk modulus, dynes/cm² × 10¹⁰

μ = rigidity (shear) modulus, dynes/cm² × 10¹⁰

λ = Lamé's constant, dynes/cm² × 10¹⁰

σ = Poisson's ratio

V_s = velocity of shear wave, m/sec

SE = Standard error of the mean

TABLE A-4: ABYSSAL PLAIN (TURBIDITE) AND ABYSSAL HILL (PELAGIC) ENVIRONMENTS:
COMPUTED ELASTIC CONSTANTS IN SEDIMENTS

Environment Sediment Type	κ		μ		λ		σ		V_s	
	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE	Avg.	SE
<u>Abyssal Plain (Turbidite)</u>										
Sandy silt	4.2572	---	0.0668	---	4.2127	---	0.492	---	201	---
Silt	3.9516	---	0.2362	---	3.7941	---	0.471	---	384	---
Clayey silt	3.0561	0.060	0.1435	0.015	2.9604	0.056	0.477	0.002	312	18
Silty clay	2.7772	0.019	0.0773	0.007	2.7245	0.020	0.486	0.001	240	11
Clay	2.7805	---	0.0483	---	2.7483	---	0.491	---	196	---
<u>Abyssal Hill (Pelagic)</u>										
Clayey silt	3.1213	---	0.1408	---	3.0274	---	0.478	---	312	---
Silty clay	3.0316	0.030	0.0795	0.007	2.9786	0.030	0.487	0.001	232	11
Clay	3.0781	0.055	0.0544	0.006	3.0419	0.058	0.491	0.001	195	12

Notes: Laboratory values: 23°C, 1 atmosphere pressure

κ = bulk modulus, dynes/cm² X 10¹⁰

μ = rigidity (shear) modulus, dynes/cm² X 10¹⁰

λ = Lamé's constant, dynes/cm² X 10¹⁰

σ = Poisson's ratio

V_s = velocity of shear wave, m/sec

SE = Standard error of the mean

TABLE A-5: CONVERSION OF DEPTH TO PRESSURE*

 $\rho = 34.0 \text{ }^{\circ}\text{‰}$, $g = 980.665 \text{ cm/sec.}^2$, (T, °C; D, meters, P, kg./cm²)

T	D	0	100	200	300	400	500	600	700	800	900
0°	0	1.03	11.31	21.59	31.87	42.17	52.46	62.76	73.06	83.37	93.69
	1000	104.01	114.33	124.66	134.99	145.33	155.67	166.02	176.37	186.73	197.09
	2000	207.45	217.82	228.20	238.58	248.96	259.35	269.74	280.14	290.54	300.95
	3000	311.36	321.78	332.20	342.62	353.05	363.49	373.93	384.37	394.82	405.27
	4000	415.73	426.19	436.65	447.12	457.60	468.08	478.56	489.05	499.54	510.04
	5000	520.54	531.05	541.56	552.07	562.59	573.11	583.64	594.17	604.71	615.25
	6000	625.79	636.34	646.90	657.46	668.02	678.59	689.16	699.73	710.31	720.89
	7000	731.48	742.07	752.67	763.27	773.88	784.49	795.10	805.72	816.34	826.97
	8000	837.60	848.23	858.87	869.51	880.16	890.81	901.47	912.13	922.79	933.46
	9000	944.13	954.81	965.49	976.17	986.86	997.55	1008.25	1018.95	1029.66	1040.37
10°	0	1.03	11.30	21.57	31.84	42.12	52.40	62.69	72.98	83.28	93.58
	1000	103.88	114.19	124.51	134.82	145.15	155.47	165.81	176.14	186.48	196.83
	2000	207.18	217.53	227.89	238.26	248.62	258.99	269.37	279.75	290.14	300.52
	3000	310.92	321.32	331.72	342.12	352.53	362.95	373.37	383.79	394.22	404.65
	4000	415.09	425.53	435.97	446.42	456.88	467.33	477.80	488.26	498.73	509.21
	5000	519.68	530.17	540.65	551.14	561.64	572.14	582.64	593.15	603.66	614.17
	6000	624.69	635.22	645.75	656.28	666.81	677.35	687.90	698.44	709.00	719.55
	7000	730.11	740.68	751.25	761.82	772.39	782.97	793.56	804.15	814.74	825.33
	8000	835.93	846.54	857.15	867.76	878.37	888.99	899.62	910.24	920.88	931.51
	9000	942.15	952.79	963.44	974.09	984.75	995.40	1006.07	1016.73	1027.40	1038.08
20°	0	1.03	11.28	21.52	31.78	42.03	52.29	62.56	72.83	83.10	93.38
	1000	103.66	113.95	124.24	134.53	144.83	155.14	165.44	175.75	186.07	196.39
	2000	206.72	217.04	227.38	237.71	248.05	258.40	268.75	279.10	289.46	299.82
	3000	310.19	320.56	330.93	341.31	351.69	362.07	372.46	382.86	393.26	403.66
	4000	414.06	424.47	434.89	445.31	455.73	466.15	476.58	487.02	497.45	507.90
	5000	518.34	528.79	539.24	549.70	560.16	570.63	581.10	591.57	602.05	612.53
	6000	623.01	633.50	643.99	654.49	664.99	675.49	686.00	696.51	707.02	717.54
	7000	728.06	738.59	749.12	759.65	770.19	780.73	791.27	801.82	812.37	822.93
	8000	833.49	844.05	854.62	865.19	875.76	886.34	896.92	907.51	918.10	928.69
	9000	939.29	949.89	960.49	971.10	981.71	992.32	1002.94	1013.56	1024.19	1034.81
30°	0	1.03	11.25	21.46	31.68	41.91	52.14	62.37	72.61	82.85	93.10
	1000	103.35	113.60	123.86	134.12	144.38	154.65	164.93	175.21	185.49	195.77
	2000	206.06	216.35	226.65	236.95	247.26	257.57	267.88	278.20	288.52	298.84
	3000	309.17	319.50	329.84	340.18	350.52	360.87	371.22	381.57	391.93	402.29
	4000	412.66	423.03	433.40	443.78	454.16	464.54	474.93	485.32	495.72	506.12
	5000	516.52	526.93	537.34	547.75	558.17	568.59	579.02	589.45	599.88	610.31
	6000	620.75	631.19	641.64	652.09	662.54	673.00	683.46	693.93	704.39	714.86
	7000	725.34	735.82	746.30	756.78	767.27	777.76	788.26	798.76	809.26	819.77
	8000	830.28	840.79	851.31	861.83	872.35	882.87	893.40	903.94	914.47	925.01
	9000	935.56	946.10	956.65	967.21	977.76	988.32	998.89	1009.45	1020.02	1030.60

*From Wilson (1959, table VIII). Wilson has similar tables for salinities 33.0 to 37.0 ppt.

TABLE A-5 (Continued)

b. $S = 35.0^\circ/\text{sec}$; $g = 980.665 \text{ cm/sec}^2$. (T, $^\circ\text{C}$; D, meters; P, kg/cm^2)

T	D	0	100	200	300	400	500	600	700	800	900
6°	0	1.03	11.32	21.61	31.90	42.20	52.50	62.81	73.12	83.44	93.76
	1000	104.09	114.42	124.75	135.09	145.44	155.79	166.15	176.50	186.87	197.24
	2000	207.61	217.99	228.37	238.76	249.15	259.55	269.95	280.35	290.77	301.18
	3000	311.60	322.02	332.45	342.88	353.32	363.76	374.21	384.66	395.12	405.58
	4000	416.04	426.51	436.99	447.46	457.95	468.43	478.92	489.42	499.92	510.42
	5000	520.93	531.45	541.96	552.49	563.01	573.54	584.08	594.62	605.16	615.71
	6000	626.26	636.82	647.38	657.95	668.52	679.09	689.67	700.25	710.84	721.43
	7000	732.03	742.63	753.23	763.84	774.45	785.07	795.69	806.31	816.94	827.58
	8000	838.22	848.86	859.50	870.15	880.81	891.47	902.13	912.80	923.47	934.14
	9000	944.82	955.51	966.20	976.89	987.58	998.28	1008.99	1019.70	1030.41	1041.13
10°	0	1.03	11.31	21.58	31.86	42.15	52.44	62.73	73.03	83.34	93.64
	1000	103.96	114.28	124.60	134.92	145.25	155.59	165.93	176.27	186.62	196.97
	2000	207.33	217.69	228.06	238.43	248.80	259.18	269.57	279.95	290.35	300.74
	3000	311.14	321.55	331.96	342.37	352.79	363.21	373.64	384.07	394.51	404.95
	4000	415.39	425.84	436.29	446.75	457.21	467.67	478.14	488.61	499.09	509.57
	5000	520.06	530.55	541.04	551.54	562.04	572.55	583.06	593.57	604.09	614.61
	6000	625.14	635.67	646.21	656.74	667.29	677.83	688.39	698.94	709.50	720.06
	7000	730.63	741.20	751.78	762.36	772.94	783.53	794.12	804.71	815.31	825.91
	8000	836.52	847.13	857.75	868.37	878.99	889.62	900.25	910.88	921.52	932.16
	9000	942.81	953.46	964.11	974.77	985.43	996.10	1006.77	1017.44	1028.12	1038.80
20°	0	10.3	11.28	21.54	31.80	42.06	52.33	62.60	72.88	83.16	93.44
	1000	103.73	114.03	124.32	134.63	144.93	155.24	165.56	175.88	186.20	196.53
	2000	206.86	217.19	227.53	237.88	248.22	258.58	268.93	279.29	289.66	300.02
	3000	310.40	320.77	331.15	341.54	351.93	362.32	372.72	383.12	393.52	403.93
	4000	414.34	424.76	435.18	445.61	456.03	466.47	476.90	487.34	497.79	508.24
	5000	518.69	529.15	539.61	550.07	560.54	571.01	581.48	591.96	602.45	612.93
	6000	623.42	633.92	644.42	654.92	665.43	675.94	686.45	696.97	707.49	718.01
	7000	728.54	739.08	749.61	760.15	770.70	781.24	791.79	802.35	812.91	823.47
	8000	834.04	844.61	855.18	865.76	876.34	886.92	897.51	908.10	918.70	929.30
	9000	939.90	950.50	961.11	971.73	982.34	992.96	1003.59	1014.22	1024.85	1035.48
30°	0	1.03	11.25	21.48	31.70	41.94	52.17	62.41	72.66	82.90	93.16
	1000	103.41	113.67	123.94	134.20	144.48	154.75	165.03	175.32	185.60	195.90
	2000	206.19	216.49	226.79	237.10	247.41	257.73	268.05	278.37	288.70	299.03
	3000	309.36	319.70	330.04	340.39	350.74	361.09	371.45	381.81	392.17	402.54
	4000	412.91	423.29	433.67	444.05	454.44	464.83	475.23	485.62	496.03	506.43
	5000	516.84	527.25	537.67	548.09	558.51	568.94	579.37	589.81	600.24	610.68
	6000	621.13	631.58	642.03	652.49	662.95	673.41	683.88	694.35	704.82	715.30
	7000	725.78	736.26	746.75	757.24	767.73	778.23	788.73	799.24	809.75	820.26
	8000	830.77	841.29	851.81	862.34	872.87	883.40	893.94	904.47	915.02	925.56
	9000	936.11	946.66	957.22	967.78	978.34	988.91	999.48	1010.05	1020.62	1031.20

*From Wilson (1959, table IX)

TABLE A-6. SAN DIEGO TROUGH: WATER PROPERTIES WITH DEPTH*

Depth (m)	Temp. (°C)	Salinity (ppt)	Velocity		Pressure (kg/cm ²)	Density (g/cc)
			(ft/sec)	(m/sec)		
0	15.08	33.46	4940.13	1505.76	1.0332	1.024833
11	14.99	33.46	4939.78	1505.65	2.1592	1.024902
31	14.80	33.46	4938.86	1505.37	4.2068	1.025033
56	12.58	33.45	4916.07	1498.42	6.7671	1.025594
66	12.26	33.57	4913.51	1497.64	7.7917	1.025794
76	11.83	33.60	4909.29	1496.36	8.8164	1.025945
91	11.31	33.62	4904.20	1494.80	10.3537	1.026125
106	11.14	33.78	4903.70	1494.65	11.8913	1.026348
126	10.60	33.84	4898.70	1493.13	13.9419	1.026583
151	10.03	33.95	4893.74	1491.62	16.5058	1.026883
201	9.63	34.15	4892.46	1491.23	21.6356	1.027335
301	8.96	34.27	4890.19	1490.53	31.9010	1.027998
401	7.48	34.25	4876.99	1486.51	42.1732	1.028676
500	6.68	34.26	4872.10	1485.02	52.3492	1.029258
650	5.73	34.34	4868.10	1483.80	67.7789	1.030148
800	4.96	34.38	4866.13	1483.20	83.2219	1.030977
949	4.36	34.43	4866.33	1483.26	98.5747	1.031784
1100	3.84	34.46	4867.58	1483.64	114.1462	1.032573

* Based on Nansen cast from R/V ARGO, Scripps Inst., 4 January 1962, at 32°40' N. Lat., 117°37' W. Long. From Miller, NURDC.

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13. ABSTRACT <p>A study of the acoustic and related properties of the sea floor in three major environments of the North Pacific: the continental terrace (shelf and slope), abyssal plain (turbidite), and abyssal hill (pelagic). Discussions cover the correction of laboratory to <i>in situ</i> values, and the prediction of <i>in situ</i> values. The methods developed are applicable to other areas and sediments.</p>		

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